THE GEOLOGY AND GEOMORPHOLOGY

OF THE

DENTON HILLS,

SOUTHERN VICTORIA LAND,

ANTARCTICA

A thesis submitted in partial fulfilment of the requirements for the Degree

of

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by N. J. Carson

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Dedication

I would like to dedicate this thesis to two very special people who passed away during this research.

Benjamin Andrew Carson

09/05/1984 – 25/04/2010

William Andrew Carson

28/01/1922 - 09/12/2009

Abstract

This research is an integrated geological and geomorphological study into the Denton Hills area. The study area is part of the foothills to the Transantarctic Mountains, which divides East and West Antarctica, allowing an opportunity to investigate glacial events from both sides. As the study area is ice-free, it has allows good examination of the bedrock geology and has preserved geomorphological features allowing them to be examined and sampled.

Comprehensive geological map and geomorphological maps have been produced, extending the knowledge into the spatial distribution of units and features. Both the geological and geomorphological maps reveal a complex history of evolution. The original geological units have been subjected to deformation and intrusion of large plutons. The geomorphological mapping shows ice has flowed in alternate direction through the valleys, and the valleys have had long periods where they have been occupied by large proglacial lakes. As the Antarctic ice sheets expanded they flowed into the valleys either from the west, the Royal Society Range draining the East Antarctic Ice Sheet or from the east, McMurdo Sound. Ice would flow from McMurdo Sound when the West Antarctic Ice Sheet expanded causing the grounding line of the ice sheet to move north through the Ross Sea.

Surface exposure dating completed during the study has correlated the timing of glacial events to global cycles. The dating confirmed the presence of the large proglacial lake during the Last Glacial Maximum in the Miers Valley, which drained about 14 ka. The Garwood Glacier has also been directly linked to the Last Glacial Maximum with a moraine forming about 22 ka. The dating has also shown that during the Last Glacial Maximum there was little fluctuation in the size of glaciers draining the East Antarctic Ice Sheet, with features being date to the onset of the Last Glacial Maximum.

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1. Introduction

The McMurdo Dry Valleys have been studied by many scientists in varied disciplines since the early explorers of Scott and Shackleton. Presently more integrated research is being conducted incorporating many different scientific disciplines. This thesis is part of a much larger biocomplexity project discussed below. The aims of this thesis are to incorporate the requirements of larger the project and to extend the knowledge into the geological and geomorphological history of the valleys. The scope of the research, the association with the larger project and an overview of the study area will be presented in this chapter.

1.1. Scope of research

This research furthers the understanding into the geological units and geomorphological features, allowing possible interpretations to be made on the evolution of the Denton Hills area. The geomorphological interpretation is supported by Surface Exposure Dating (SED) of samples (collected during the field season), which allow age constraints to be assigned to geomorphological features.

The scope of the study can be split into three main aims:

1. To create maps showing the spatial distribution of the geological units and geomorphological features within the Denton Hills study area. This can be used as an input into the nzTABS models to confine the environmental factors into predicting biocomplexity.

2. To present possible scenarios for evolution of the valleys using the maps, descriptions of the geological units and geomorphological features. As part of the study surface exposure dating (SED) was used to place age constrains on the formation of the geomorphological surface and/or last possible ice cover.

3. To determine the timing of glaciations and deglaciation in the area by using SED dating of geomorphological features directly related to glaciers.

The research has improved the knowledge into the spatial distribution of the geological units and geomorphological features, which are used for a larger international Antarctic project. This larger project is an International Polar Year (IPY) project; New Zealand Terrestrial Antarctic Biocomplexity Survey (nzTABS), which is creating a model to predict the distribution of plants and organisms by using landscape features such as bedrock geology and geomorphic features. The model will be used to investigate the physical controls on the diversity, quantity and distribution of the flora and fauna throughout the extreme environment of the Dry Valley ecosystem.

1.2. Association with IPY project (nzTABS)

This research is associated with the joint University of Waikato (UoW) and University of Canterbury (UoC) IPY project, nzTABS, "Predicting Biocomplexity in Dry Valley Ecosystem". The aim of the nzTABS project is to create a model that allows the biodiversity and biological activity to be predicted by using the landscape and environmental factors. The factors which were investigated were; altitude, aspect, slope, geology and geomorphology as possible factors into the distribution of biological activity.

The study area selected for the project was the Denton Hills, southern McMurdo Dry Valleys, Southern Victoria Land, Antarctica. The area was chosen because of on-going biological studies undertaken by members of the team within the Miers and Garwood Valleys. The project then extended the research area to include the smaller Marshall Valley situated between the two other valleys. Some areas of the study area had been studied previously for biological, hydrological, geological and geomorphological studies, however there had been no individual study covering the entire area or investigating all the environmental factors.

This thesis is contributing complete geological and geomorphological maps of the area and also providing some SED dating allowing interpretations into the ages of surfaces. The age of surface can determine the period of time the surface has been stable, e.g. ice free and/or material has stopped being deposited. These ages can be used as an indication about the length of time biological activity has been able to colonise the surface features.

For geological and geomorphological study, the nzTABS project allowed two weeks during early November 2008, to complete the field mapping, observations and collecting of samples for Surface Exposure Dating (SED). During the two weeks, two main field camps were setup with five sub-camps positioned around the field area, shown in Appendix 1. During the two weeks most of the time was spent in the Garwood Valley, staying at the main camp situated between the Garwood Glacier and Lake Colleen. We moved throughout the valleys, mapping and photographing the geological units and geomorphological features.

1.3. Methodology

Geological and geomorphological maps: maps were first drafted using information from previous studies and drawn onto ALOS (Advanced Land Observation Satellite) images and LIDAR (Light Detection and Ranging) data. These draft maps was then printed onto laminated base maps and taken into the field during the November field season. Over the two week field season the boundaries, lithologies and features were ground truthed, modifications were drawn directly onto the base maps. The final maps were drawn by using CorelDraw and ArcGIS mapping digital software.

Surface Exposure Dating (SED): use of *in situ* ^{10}Be and ^{26}Al cosmogenic radionuclide dating on rock directly associated with glacial events. SED is a dating method that indicates the period of time in which a rock surface has been exposed to cosmic rays from space. The cosmic rays interact with atoms contained within quartz, this interaction creates radioactive isotopes. The ratio of the isotopes can provide an estimated period of exposure. The process to remove the quartz from the rock samples was conducted in the Cosmogenic Preparation Laboratory, Department of Geological Sciences, University of Canterbury (UoC). After the quartz was refined it was analysed at Australian Nuclear Science Technology Organisation (ANSTO) on the ANTARES-AMS accelerator (Explained in further detail in Chapter 5).

By combining previous studies, maps, field observations and new data, interpretations into the evolution the valleys were formed. The evolution from the initial geological units and uplift of the Transantarctic Mountains (TAMs) through to creation of the recent geomorphic landforms was created. By including the SED data to the evolution scenarios, age constraints were assigned to the geomorphological features, which were used to better understand the timing of formation of these features and the chronology of the ice movement throughout the valleys.

1.4. Study Area

The study area is approximately 220 km^2 area of ice free land, in an area known as Denton Hills. Denton Hills are located south of the main McMurdo Dry Valleys area (Taylor, Wright and Victoria Valleys) in Southern Victoria Land, Antarctica. The study area is between 163°39’000E - 164°28’000E and 78°00’000S - 78°08’000S as shown in Figure 1.1 & 1.2. The study area incorporates three valleys (Miers, Marshall and Garwood Valleys) which have flat valley floors with steep valley sides rising up to peaks over 1000 m. The Miers Valley has an asymmetrical profile with the southern wall being at a shallower slope than the northern. These mountains are known as the southern foothills, to the much larger and spectacular Royal Society Range further to the west. To the west the study area is bound by glaciers draining the Royal Society Range and the east by McMurdo Sound. The coastline along this section of McMurdo Sound is ice covered, from ice getting discharged from the Koettlitz Glacier which flows into the McMurdo Ice Shelf (Figure 1.1).

Each valley has different features including; profiles, lengths, lakes and glaciers;

The Miers Valley is the longest approximately 15 km long and 5 km wide in the centre. The Miers Valley contains two glaciers (Miers and Adams Glaciers) which flow in from the west around a peak known as Holiday Peak. The wide flat floor of the valley can be split into two basins (east and west). Presently within the west basin, Lake Miers is fed by meltwater from the Miers and Adams Glaciers and winter snow drifts. The eastern basin has a river flowing through it from Lake Miers to McMurdo Sound.

The Marshall Valley is a smaller valley approximately 12.5 km long and 4 km wide. At the head of the valley is a semi-circle escarpment which the Rivard Glacier occupies. The Marshall Valley does not connect into any outlet glaciers of the EAIS therefore the Rivard Glacier is typified as an alpine style. Unlike the other valleys the Marshall Valley does not show a flat base profile but still has steep sided valley walls. A well-defined stream flows through the valley from the face of the glacier to the mouth of the valley and into McMurdo Sound.

The Garwood Valley is approximately 13 km and 6 km wide. The Joyce Glacier is situated at the head of the valley flowing in from the west as part of the EAIS drainage system. Meltwater from both the Joyce Glacier and from Buddha Lake (south of the Joyce Glacier) flow into Lake Colleen situated at the western end of the valley in-between the Joyce and Garwood Glaciers. The Garwood Glacier is a much smaller alpine glacier, fed by the surrounding mountains (e.g. Williams Peak). A meltwater stream transports meltwater and sediment from Lake Colleen around the snout of the Garwood Glacier, down the valley to McMurdo Sound.

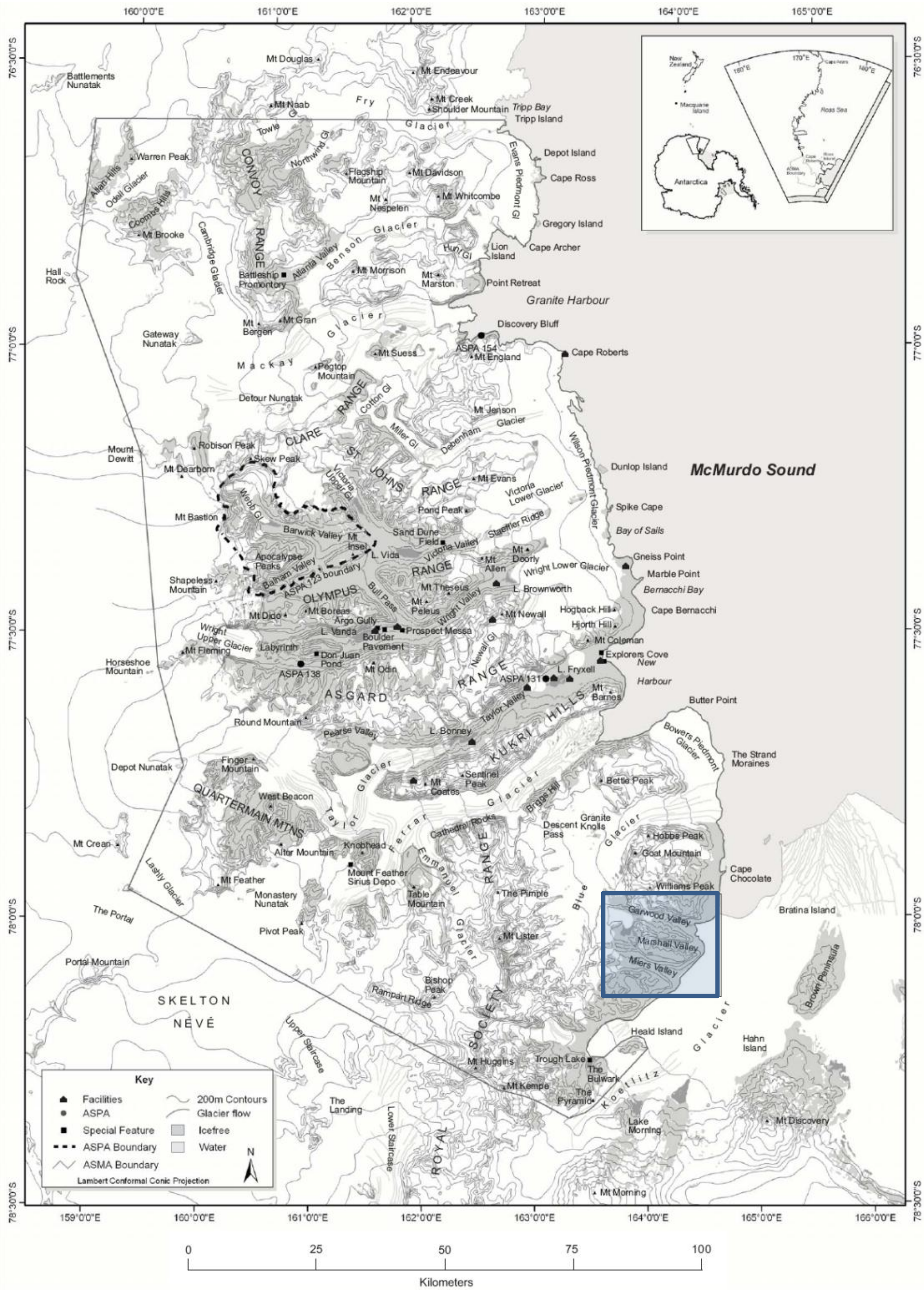

Figure 1.1: Location map of the McMurdo Dry Valleys, Southern Victoria, Antarctica. The maps indicate where the satellite photo (Figure 1.2) is located. The Denton Hills field area is shown by the square outline. (Image from Antarctica New Zealand)

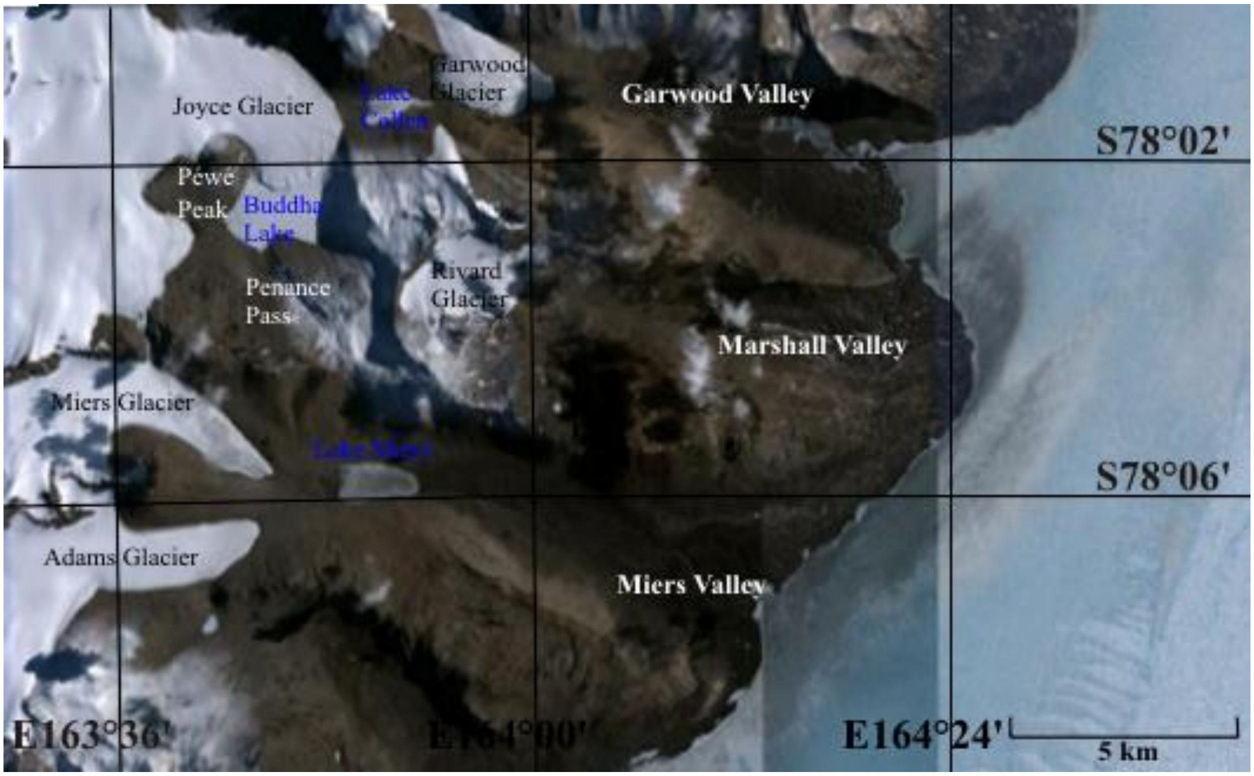

Figure 1.2: Close up Google-Earth image of the Denton Hills field area.

1.5. Regional Geological Setting

The study area lies in the foothills of the Royal Society Range, a 70 km section of the Transantarctic Mountains (TAMs). The TAMs are a long and high mountain belt that transects the continent over 3000 km, forming the boundary between East Antarctica and West Antarctica. The Royal Society Range is marked by the spectacular escarpment (Figure 1.3), with Mt Lister reaching up to 4025 m elevation. The escarpment is suggested to be the rift margin of the West Antarctic Rift system (Gleadow & Fitzgerald 1987; Fitzgerald 1992; Jones 1996 and Sugden et al., 1999). The Royal Society Range drains east into the Blue Glacier which feeds the Joyce, Miers and Adams Glaciers which flow into the study area.

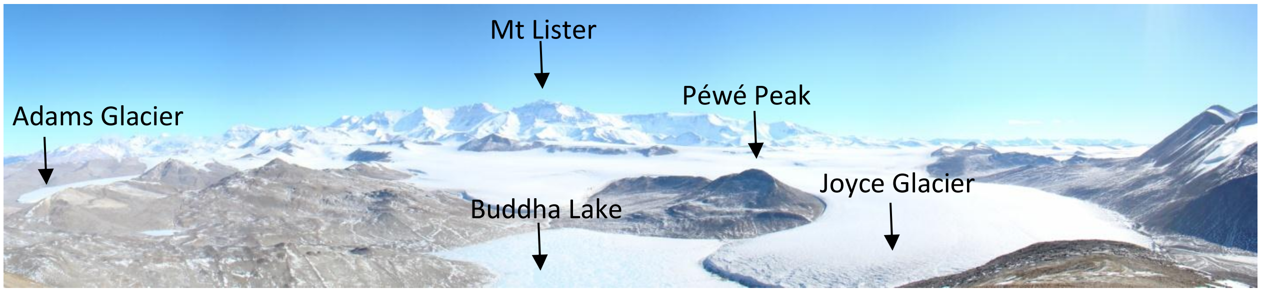

Figure 1.3: The spectacular escarpment of the Royal Society Range, with Mr Lister (4025 m) being the highest point. The Joyce Glacier flows around Péwé Peak and Buddha Lake in the foreground. To the extreme left the Adam Glacier can just be seen flowing into the Miers Valley.

The basement geology of the Denton Hills and TAMs is comprised of Pre-Cambrian to Ordovician metasediments with intruded Palaeozoic granitoid plutons (Skinner, 1983; Findlay et al., 1984; Allibone et al., 1993). The metasediments are comprised of biotite gneiss and schist with a thick band of marble. Throughout the area, there are differing granitoid bodies in both chemistry and deformation, indicating that the granitoids have been emplaced at various stages during the period of the west dipping subduction along the eastern craton margin (Jones, 1996).

As the metasediments were uplifted, erosion occurred removing some of Ordovician sediments. This unconformity throughout the McMurdo Dry Valley area is known as the Kukri Peneplain (Gunn and Warren, 1962). The Kukri Peneplain was flat lying erosion surface which then had Devonian to Jurassic age sediments deposited upon it, these sediments are known as the Beacon Supergroup (Findlay et al., 1984). The Beacon Supergroup is comprised of siltstones, sandstones and conglomerates, which presently dip gently to the west (away from the Ross Sea) (Sugden et al., 1999).

The formation of the TAMs was started by a rifting process between the East Antarctic craton and the crustal blocks of West Antarctica, estimated to have started in the Jurassic (Elliot, 1975). The uplift of the TAMs starting about 55 Ma continued into the Oligocene where most of the uplift occurred (Gleadow & Fitzgerald, 1987; Barrett et al., 1991; Fitzgerald, 1992). The TAMs were created by several blocks being uplifted along the rift margin, created a towering mountain belt with peaks rising over 4000 m. The Royal Society Range is one of these blocks, being about

70 km long. During the rifting and uplift of the TAMs, subsidence occurred in the Ross Sea embayment (Gleadow & Fitzgerald, 1987).

Since the uplift of the TAMs, the region has undergone erosion to form the landscape observed today. The formation of the Dry Valleys were initially started by alluvial processes (Sugden et al., 1999) which created the west to east draining valleys from the TAMs. Over the history of the valleys, it is apparent there have been glacial advances from both the west (Royal Society Range) and east (Ross Sea). Some of these advances were large enough to fully cover the study area (Denton et al., 1983). The sequence of reoccurring advances and retreats have glacially eroded the valleys; widening, deepening and creating the steep sided valleys observed presently.

The McMurdo Dry Valleys is a significant proportion of the total ice-free land (only about 2% of the landmass is ice-free) currently found in Antarctica. For this reason and that the area is within 150 km of scientific bases of New Zealand, United States of America and Italy, the McMurdo Dry Valleys have been the focus of many geological and geomorphology studies. This allows greater initial understanding in the regional geological formation discussed in further detail in the following chapter.

1.6. Antarctic Climate

Antarctica is an isolated landmass situated at the southern end of the globe, this isolation at the pole allows the circumpolar vortex to form and cool the continent. The Antarctic continent is known as a frozen dessert having cold temperatures (mean annual temperature in the Dry Valleys -14.8 to -30.0 °C), little precipitation (about 10 mm water equivalent) and very strong wind (up to 320 km/hr) (Doran et al. 2002).

The study area is located in southern McMurdo Dry Valleys, Southern Victoria Land, which are ice-free due to the low humidity/precipitation and katabatic winds. The McMurdo Dry Valleys have an estimated annual temperature of -14.8 to -30.0 ° C for the coastal locations (Doran et al., 2002). These valleys are hyperarid with precipitation estimates of about 10 mm (water equivalent) per year (Thompson, 1973). Katabatic winds are formed by gravity pulling down the dense cold air sitting on the East Antarctic Ice Sheet (EAIS). The cold dense air then flows off the high EAIS plateau through the valleys at speeds exceeding 130 km/hr (Nylen et al., 2004), which during winter this can increase the ambient temperature to just below freezing.

During the summer season the warmer temperatures, which can exceed +5°C (Sugden et al., 1999) and the constant sunlight allows meltwater to be produced and flow throughout the valleys through the network of streams and lakes. The warmer temperatures also allow the partial melt of some of the ice cover of the lakes. The partial melting allows a moat to form around the edges of the lakes, and smaller ponds may completely lose their ice cover. The movement of meltwater transports sediment throughout the valleys, creating landforms by deposition or erosion creating different geomorphological landforms.

1.7. Styles of Antarctic Glaciation

1.7.1. Styles of Glaciation

The relationship between the volume of ice in Antarctica and the volume of ice either in the Northern Hemisphere or globally has been debated intensely since the discovery of the Antarctic continent. Since the discovery by the early explorers, it was recognised that Antarctica has had larger ice cover in the past (Scott, 1905; David & Priestly, 1914; Taylor, 1914; Mellor, 1959; Voronov, 1960; Aughenbaugh, 1961).

Many of the early scientists suggested that glaciation started in Antarctica during the Tertiary (Nordenskjöld, 1911; Wright & Priestly, 1922; Priestly, 1923; Gould, 1939, 1940; Nichols, 1963) and that a permanent ice sheet may have been present in Antarctica before any permanent ice sheet in the Northern Hemisphere had appeared.

From early studies three hypotheses of how Antarctica was glaciated and a chronology of these glaciations have been produced;

1. A catastrophic surge ice sheet (Wilson, 1964). Wilson (1964) suggested that the Antarctic continent accumulated ice during interglacial periods. The thickened ice volume becomes

unstable as the pressure of the ice creates temperatures less than freezing at its base. The melting creates meltwater at the ice/bedrock interface (Robin, 1955). The presence of meltwater speeds up the flow of ice off the continent, creating large ice shelves, causing an increased albedo affect, cooling the global temperatures. This state cannot continue as the accumulation rate of ice on the continent cannot support the increased speed of the flow of ice off the continent, this causes the rapid disintegration of the ice shelf and reducing the albedo.

2. In-phase reaction to sea-level change induced by the Northern Hemisphere (Hollin, 1965). This hypothesis suggests that the size of the ice sheet and glaciers in Antarctica are directly related to sea-level and that the sea-level is related to the ice volume in the Northern Hemisphere. Increased ice volume within the Northern Hemisphere causes the sea-level to fall, which displaces the Antarctic ice grounding line downwards/northwards allowing the ice to advance. When the Northern Hemisphere ice volume decreases the sea level rises creating the grounding line to move up/southwards reducing the size of the Antarctic ice cover.

3. 'Out-of-phase' fluctuations, due to greater snowfall during interglacial periods (Scott, 1905; Markov, 1969). This hypothesis suggested due to the extreme cold temperatures, the ice volume is directly related to the amount of precipitation. The mean annual temperature for Antarctica is significantly below 0 °C, during the warmer interglacial periods it would still be below freezing for most of the year, but precipitation would increase. Warmer atmosphere would allow greater evaporation and transport distance of the moisture onto the continent. Warmer oceans would reduce the extent of sea ice allowing open water to be evaporated closer to the continent. The increase precipitation would increase the volume of the ice on the continent but would restrict the ice volume in contact with the oceans, due to increased melting.

Later studies (Denton & Armstrong, 1968; Denton et al., 1970; Denton et al., 1989; Marchant et al., 1994) suggested that both in-phase and 'out-of-phase' glaciation was occurring in Antarctica.

Most of the continent ice sheet and drainage glaciers appear to be in-phase with global glaciations, however outlet glaciers which terminate on land appear to display an 'out-of-phase' response. Marchant et al., (1994) used dating methods on glacial deposits associated with the Taylor Glacier in the Arena Valley, concluding that the Taylor Glacier was responding 'out-of-phase' and advancing during interglacial periods. As these studies have suggested that there are two differing chronology of glaciations on the Antarctic continent, I will continue by discussing them separately.

1.7.1.1. Glaciers terminating in the ocean

Most of the Antarctic's ice/glaciers appear to respond to an 'in-phase' glacial cycle, reacting with global glacial and interglacial periods (Hollin, 1962; Denton & Armstrong, 1968; Denton et al., 1989). The ice sheets (EAIS and WAIS) are drained by glaciers flowing down slope to the sea, draining ice off the land. With the limited amount of ice-free land in Antarctica, much of the ice and glaciers terminate into the ocean. Glaciers both erode and deposit material leaving behind markers of the previous height and volume. By dating these features correlations between the timing of glaciation can be compared to the Northern Hemisphere and sea-level changes.

Denton & Armstrong (1968) studied the chronology of ice within the McMurdo Sound. The evidence suggests that the ice from the Ross Sea (WAIS) correlated to global glaciation. The evidence shows that ice from the Ross Sea has expanded into the mouths of the McMurdo Dry Valleys at least four times. The timing of these expansions correlates with the global glacial periods and dissipation of the ice with a rapid sea-level rise and also the melting of ice sheet in the Northern Hemisphere.

Further proof for this concept came with a more detailed study into the glacial history of the Ross Sea Embayment (Denton et al., 1989). The study investigated the chronology of glaciers draining the EAIS terminating in the ocean/ice shelf. The study concluded that the Reedy, Beardmore, Hatherton and Darwin Glaciers behaved in association with global advances and correlated to the expansion of ice through the Ross Sea Embayment.

1.7.1.2. Glaciers terminating on land

The timing of fluctuations of EAIS outlet glaciers which terminate on land has been suggested to be ‘out-of-phase’ of global glacial periods (Denton & Armstrong, 1968; Denton et al., 1989; Marchant et al., 1994). ‘Out-of-phase’ glaciation means the ice expands and advances during global interglacial periods and retreats during glacial periods. There are only a few areas in Antarctica that could allow this process to occur as the glaciers have to terminate on land, meaning the McMurdo Dry Valleys are a great area to investigate this phenomenon.

Denton & Armstrong (1968) suggested that only a few Antarctic glaciers behaviour ‘out-of-phase’ of the global glacial cycle. This theory was created from evidence that showed the Taylor Glacier had been considerably smaller during the same period that the ice from the Ross Sea had pushed up into the mouth of the Taylor Valley. Both the Taylor and Wright Glaciers both expanded once the ice from the Ross Sea had retreated. Alpine glaciers within the Taylor Valley and the Royal Society Range have also been suggested to be fluctuating ‘out-of-phase’ to the Ross Sea Embayment fluctuations.

This was further expanded when the history of the ice with the Ross Sea Embayment was investigated by Denton et al., 1989. The studies looked at EAIS drainage glaciers, and compared the relationship between the advances from the glaciers and the ice from the Ross Sea. To prove the ‘out-of-phase’ glaciation of the glaciers that terminate on land, the Taylor Valley was used, with the Taylor Glacier flowing in from the west and the Ross Sea to the east. The glacial features indicate an overlapping relationship suggesting that the two glaciers would have occupied the valley at different times.

This hypothesis was further extended by Marchant et al., (1994) investigating the chronology of glacial movement of the Taylor Glacier in the Arena Valley. Evidence from the glacial geomorphology provided evidence of five advances. A range of dating methods; soil data, SED dating (Surface Exposure Dating) and weathering data, were used to create a chronology for the advances.

The interpretation for ‘out-of-phase’ fluctuations is the precipitation is the main cause for the glaciers to advance. During global interglacial periods, sea ice surrounding the continent is reduced due to the warmer climate. The lack of sea ice surrounding the continent allows the open ocean to be closer to the landmass and therefore increased precipitation occurs in the catchment areas for the glaciers.

1.7.2. Alpine Glaciers

The term alpine glacier refers to small local glaciers formed from small mountain catchments and flow down slope. They do not receive any ice from other sources (e.g. ice sheets). As the glaciers are isolated from major ice sources, they are only fed by local precipitation. This reliance on local precipitation suggests they would behaviour ‘out-of-phase’, provided they are in an area where the temperature is below freezing point for most of the year. It is suggested the during interglacial periods precipitation increases as the proximity to an open water source is reduced. This suggests that the alpine glaciers should correlate to the ‘out-of-phase’ glaciation model (Denton et al., 1989).

2. Previous Studies

2.1. Introduction

The Denton Hills study area combines three valleys (Garwood, Marshall and Miers Valleys) located at the southern end of the McMurdo Dry Valleys, Southern Victoria Land, Antarctica. As these valleys are smaller than Taylor and Wright Valleys, to the north, the area has been studied in less detail. This chapter summarises the previous research in the Denton Hills area describing the Geological units and geomorphological features have been mapped on Appendix Sheet 1 & 2.

Cosmogenic nuclide dating (SED) is a dating technique that allows ages to be placed for the formation of surfaces. Discussed here is an overview of previous studies using cosmogenic nuclide dating within the McMurdo Dry Valleys, the method is discussed in a later chapter.

2.2. Geological Studies

2.2.1. Early Explorers

The geology of the area has been studied since the first expeditions down to Antarctica, with both Scott and Shackleton having geologists within their science parties. The initial studies were completed around Ross Island but they soon extended into the Transantarctic Mountains and into the McMurdo Dry Valleys. These studies were conducted by Ferrar, (1907); Prior, (1907); David & Priestley, (1914); Mawson, (1916); Taylor (1922); Smith & Debenham, (1921); Smith, (1924). These early studies established the major rock types, metasediments, granites, sediments and dyke/sills. Ferrar (1907) recognised that the Royal Society Range is comprised of granites, Beacon sediments and diorite sills, with considerable proportion of the Southern Foothills being comprised of metasediments. Farrar (1907) and Taylor (1922) produced geological maps of the area but were in limited detail and were drawn in large scales e.g. 1:500,000 (Taylor, 1922).

2.2.2. IGY (International Geophysical Year) Research

After the early explorers there was a long period where the continent was not explored and little scientific research was completed, until the International Geophysical Year (IGY) 1957/58. The IGY was the start of a large increase in Antarctic scientific research with 12 nations partaking, building 45 research bases (Wilson & Crary, 1961). During the IGY both New Zealand and United States of America had scientific research parties in the McMurdo Dry Valleys and/or the Transantarctic Mountains.

The first major geological studies were completed by members of the Commonwealth Trans-Antarctic Expedition (1955-58). There were seven parties spread out along the Transantarctic Mountains, using sledges either pulled by dog teams or manhauled, reducing the size of the area which could be covered and also the size of geological specimens. This work was collated by Gunn & Warren, (1962) publishing the first major geological study which included a map on a scale of 1:250,000 of the McMurdo Dry Valleys. The parties covered much of the southern Victoria Land, describing the geological units and mapping the areas in which they visited. No party entered the Denton Hills area, although the units within the area were assumed to be the same geologic units as found throughout the McMurdo Dry Valleys. The study identified the oldest unit in the McMurdo Dry Valleys to be the metasediments, named 'Skelton Group' which included marble, schist and gneiss. The study also mapped all intrusive rocks older than the Kukri Peneplain (a regional marker) as the same complex named the 'Granite Harbour Intrusive Complex' which was mapped as a significant proportion of the mapped area. The geological map showed in the Denton Hills area as containing marble along the ridgelines and Quaternary deposits within the valleys, the 'Granite Harbour Intrusive Complex' borders the study area and was suggested to extend into the area.

The first geological map including the Denton Hills study area was by Blank et al., (1963). This exploration party of five men from Victoria University of Wellington studied the geology in the area between Blue Glacier (77° 50') and Koettlitz Glacier (78° 30').

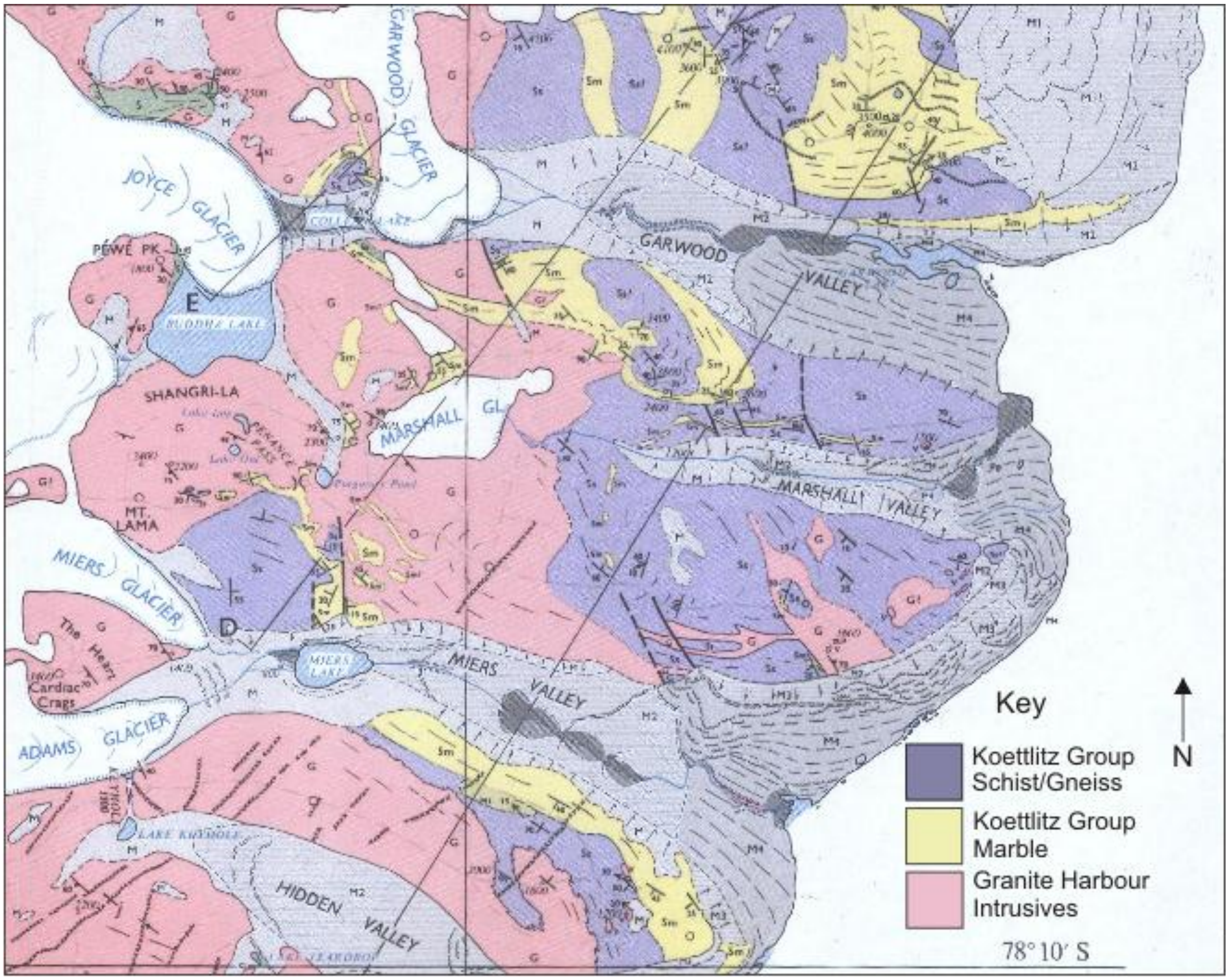

Figure 2.1: The Denton Hills section of Blank et al., (1963) geological map. The map generalises the lithology grouping schist and gneiss in one group and marbles in another. The map shows how the western side of the field area is dominated by the 'Granite Harbour Intrusives' and the eastern by the Koettlitz Group.

The lithology identifications were based on field observations and the examination of 150 thin sections. Blank et al., (1963) showed that in the east, the Denton Hills area is dominated by the metasediments (‘Koettlitz Group’) containing five different formations comprised of schist, gneiss and marbles.

1. Hobbs Formation – distinctive dark schist thought to be the oldest rocks in this area. Large clasts of pebble to cobble sizes of granite, quartzite and calc-silicate are found through the matrix of the rock, and have been stretched at least twice there original size

2. Salmon Marble – can appear through the field up in thickness over 2500 m of pink/grey coarse grained marble. Bedding is marked by graphitic layers, and cylindrical siliceous objects were found within the marble thought to be archaeocyathids.

3. Garwood Lake Formation – a 600 m thick formation of fine grained leucocratic biotite, augen- and banded gneiss, quartz-biotite schist, and a Hobbs like conglomerate quartz-tremolite schist.

4. Miers Marble – blue/grey to cream/white marble with highly contorted schist rolled up within it, in some areas has the appearance of the Salmon Marble.

5. Marshall Formation – found in the eastern end of the valleys, starting at the coastline. It is suggested to be approximately 450 m of interbedded biotite schist, thin marble, banded quartzite, Hobbs like granulite and calc-silicate granulite.

In the west the area is dominated by a granitoid intrusive ('Granite Harbour Intrusive Complex') which show evidence for several different generations. They also noted that the entire area was scattered with basaltic dikes, which could be easily seen contrasting against the bedrock geology. They also identified a 'recent' basaltic volcanic deposit at the end of the Walcott Valley, just south of Miers Valley. Blank et al., (1963) also mapped the limits of the 'Ross Sea Drifts' which are found as dark volcanic rock deposited in the eastern end on the valley floors.

In 1964, Grindley & Warren published all the geological research work that had been completed during the IGY along the Transantarctic Mountains, to correlate and make recommendations for naming units. This was requested by the New Zealand Committee for SCAR (Scientific Committee of Antarctic Research), and was prepared by the Geological Society of New Zealand (GSNZ). Grindley & Warren (1964) correlate the old metasediments known as the 'Skelton Group' and the 'Koettlitz Group' with the Robertson Bay and Berg Groups in the north and the Byrd and Beardmore Groups in the south, naming them the Koettlitz Group. The 'Granite Harbour Intrusives' are identified throughout the area from Terra Nova Bay to the Shackleton

Glacier. Quaternary volcanics associated with the formation of Ross Island and other dominate landmarks, are found outcropping throughout the study area. The authors also mention how the area has been highly affected by Quaternary glaciation and fluvial processes, commenting on the studies of Péwé (1958a, b, 1959, 1960, 1962).

2.2.3. Research post IGY

After the major scientific advances in research during the IGY, geological studies continued mainly from national programs. New Zealand and United States of America continued to study the McMurdo Dry Valley area. The research became more detailed investigating individual unit/formations, which also allowed interpretations into the deformation history to be formed by studying both mega and micro structures within units. This deformation history is discussed in Appendix 2.

In 1972 Lopatin investigated the basement complex of the McMurdo 'Oasis' (Dry Valleys), working mainly on the Taylor and Wright Valleys. The Denton Hills study area was investigated in less detail, but Lopatin (1972) concluded that many of the same lithologies and units are observed in the north. The metasediments were studied between the Blue and Koettlitz Glaciers and were alleged to be a mix of; biotite-amphibole-schist, quartz-diopside-amphibole- and quartz-amphibole-schist, biotite-muscovite- and quartz-muscovite-schist and thick beds of marble (Lopatin, 1972).

A detailed geological study of the area between the Renegar and Blue Glaciers was conducted by Findlay (1978), which investigated the metamorphosed units and completed a detailed structural investigation of the area (Appendix 2).

Mortimer (1981) investigated the basement geology between the Salmon and Miers Valleys, which establish several contradicting interruptions than that of the earlier study by Blank et al. (1963). Younging indicators which had been identified by Blank et al. (1963) within the Koettlitz Group could not be found by Mortimer (1981), suggesting the stratigraphy of the Koettlitz Group uncertain. Studies into the Salmon and Miers Marbles, suggested that Findlay (1978) was correct by suggesting that the two type of marble that Blank et al., (1963) had identified and mapped, appeared to be the same marble and could not find any reason to differentiate between them. By tracing beds from one valley to the next, they proved that the same beds within the Marshall Valley are present in the Garwood Valley, this proved that the Marshall and Garwood Formations suggested by Blank et al. (1963) were the same formation.

The Koettlitz Group was the main focus of the New Zealand Geological Survey and New Zealand Antarctic Division between 1977 and 1981. This work was summaries by Findlay et al. (1984), which investigated the lithostratigraphy and structure. The Koettlitz Group contains several formations and has undergone several phases of deformation as shown by the previous studies. Findlay et al. (1984) split the Koettlitz Group into three formation; Marshall, Salmon and Hobbs Formation, reducing from Blank et al. (1963) initial five formations.

- The Marshall Formation consists of biotite schist, thin marbles and paragniess.

- The Salmon Formation consists of two members; Heald and Dismal Members. The Heald Member is well developed white to cream marble with about 10 cm thick rusty red layers of pyrite, muscovite and diopside. Within the member at the north end there is a 10-20 m thick intercalations of biotite schist. The Dismal Member is a cream-orange to grey-black marble with interbeds of hornblende and biotite.

- The Hobbs a Formation consists of four members; Radian Schist, Con-Rod Hills, Rucker and Meserve Members. The Radian Schist Member is purple-brown biotite schist which follows on from the Dismal Member of the Salmon Formation. The Radian Schist grades into Con-Rod Hills Member, which comprises of fine grained tremolite/actinolite matrix with rare pebble and cobbles clasts of granite and marble. Con-Rod Member grades into the Rucker Member, a medium grained tremolite/actinolite amphibolite containing rare pebbles of quartz, granite and phelite. Finally the sequence grades into the Meserve Member comprised of biotite schist.

The structural interpretation was the same as that published earlier by Findlay (1978), refer to Appendix 2.

2.2.4. Otago University Geology Department Antarctic Research

During the late 1980's and early 1990's Otago University had several geological projects focusing on geological units and structure of this area of Southern Victoria Land. Many of these projects were MSc projects supervised by Dr. David Craw, Dr. Alan Copper and Dr. Richard Norris.

- Koettlitz Group – Allibone (1988)

Allibone (1988) completed a detailed study into the lithology and structure of the metasediments which form the Koettlitz Group. The Koettlitz Group is comprised of marbles, schists and minor parts of pelite. Marble is the greatest component within the group; this marble has been given the name the Salmon Marble, the remaining proportion is dominated by migmatitic quartzofeldspathic and psammitic schist. Allibone (1988) suggested only two phases of deformation have affected the Koettlitz Group; firstly isoclinal folds and secondly tight, upright, shallow plunging north northwest trending folds.

- Gneiss Geology – Cox (1989)

Cox (1989) studied in detail the foliated 'Granite Harbour Intrusive' and the host rocks (Koettlitz Group) throughout the Wright Valley, producing the deformation history. The granitoid pluton is elongated northwest-southeast following along the trend of the Transantarctic Mountains. Both lineations within the pluton and host metasediments were found to follow a parallel trend to the elongation of pluton. This trend suggests a regional tectonic event controlled the emplacement of the granitoid pluton. Study into the deformation within the Koettlitz Group confirmed the previous hypotheses (Findlay, 1978; 1984; Allibone, 1988) that the group has undergone three phases of deformation. This deformation was accompanied by amphibolite facies metamorphism. Cox (1989) completed geothermometry on the metaphoric rocks to identify a possible 'peak' metamorphic temperature. By using the Fe-Mg garnet-biotite geothermometry a temperature of 680° ± 40° C was suggested to be the ‘peak’ metamorphism temperature. The first two deformation events were syn-migmatitic deforming in a plastic state (Cox, 1989). There are also some granitoids identified by Cox (1989) to have been emplaced after the deformation, implying they were younger than the deformation.

- Granite Harbour Intrusives, Antarctica – Smillie (1989)

This was a detailed study into the granitoids, which suggested there are two different suites (Dry Valleys 1 Suite (DV1) and Dry Valley 2 Suite (DV2)) separated by differing mineralogies (Figure 2.2), textures and intrusive patterns.

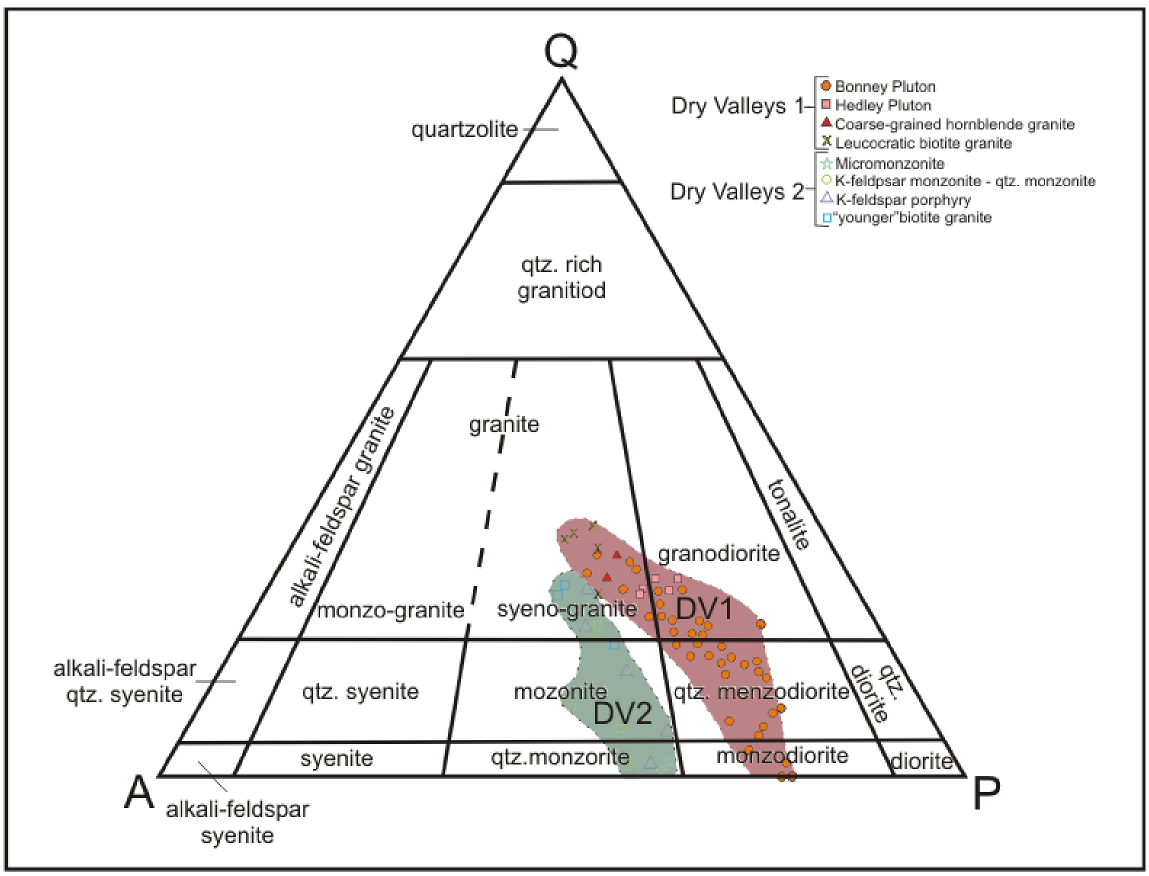

Figure 2.2 QAP (Quartz, Alkali-feldspar Plagioclase-feldspar) diagram from Smillie (1989). This shows the differing chemical characteristics between DV1 and DV2.

DV1 has a calc-alkaline, I-type chemistry, suggesting a continental arc environment for formation (Smillie, 1989). Within the DV1 the Bonney Pluton has a regional flow foliation parallel to the long axis of the pluton, which could indicate a major shear zone, likely to be associated with a plate boundary near a subduction zone. Evidence shows that there was partial melting of the crust as the suite was emplaced. In northern Victoria Land it is consider that the and 'Granite Harbour Intrusives' have a continental arc environment (Borg et al., 1984) suggesting the subduction zone would extend far north, possibly along the entire length of the Transantarctic Mountains within Victoria Land.

DV2 has an alkali-calcic, I-type chemistry with high K_2O and low CaO contents. This chemistry suggests a post major tectonic regime and unlikely to be related to the subduction process (Smillie, 1989). The subduction would have ceased allowing upwelling to melt the crust creating large volumes of granite. The greater number of dike swarms suggest a period of extension occurred allowing the material to infill.

- Deformation South Victoria Land – Cook (1992)

This investigation into the deformation of the basement rock in Southern Victoria Land, showed the regional shear zone of orientation NW-SE, was initiated and active throughout deformation which occurred during the emplacement of the syntectonic pluton intrusion ('Granite Harbour Intrusive Complex'). The deformation has also extensively changed the original structure and chemistry of the Koettlitz Group, both by the re-crystallisation of minerals and element exchange with the neighbouring mega-gabbro and granitoids. Agreement with the hypothesis for three stages of deformation (Findlay, 1978; 1984; Allibone, 1988; Cox, 1989) of the Koettlitz Group was further strengthen by this study.

- Dismal Geology – Worley (1992)

Worley (1992) further extended the study into the granitoids using an integrated study of field observation and geochemical analysis. This study extended the subdivision of the 'Granite Harbour Intrusive' by Smillie (1989) into DV1 and DV2. Many of the granitoids have similar appearance in the field and therefore had been previously mapped together by Gunn & Warren (1962), Blank et al., (1963), Grindley & Warren (1964) and Smillie, (1989). Worley (1992) by using the geochemical data identified at least six intrusions in the Miers Valley; Bonney Pluton, Buddha Diorite, Lama Pluton, Rivard Diorite, Miers Granite and Biotite Granite. Worley's sketch map is shown in Figure 2.3.

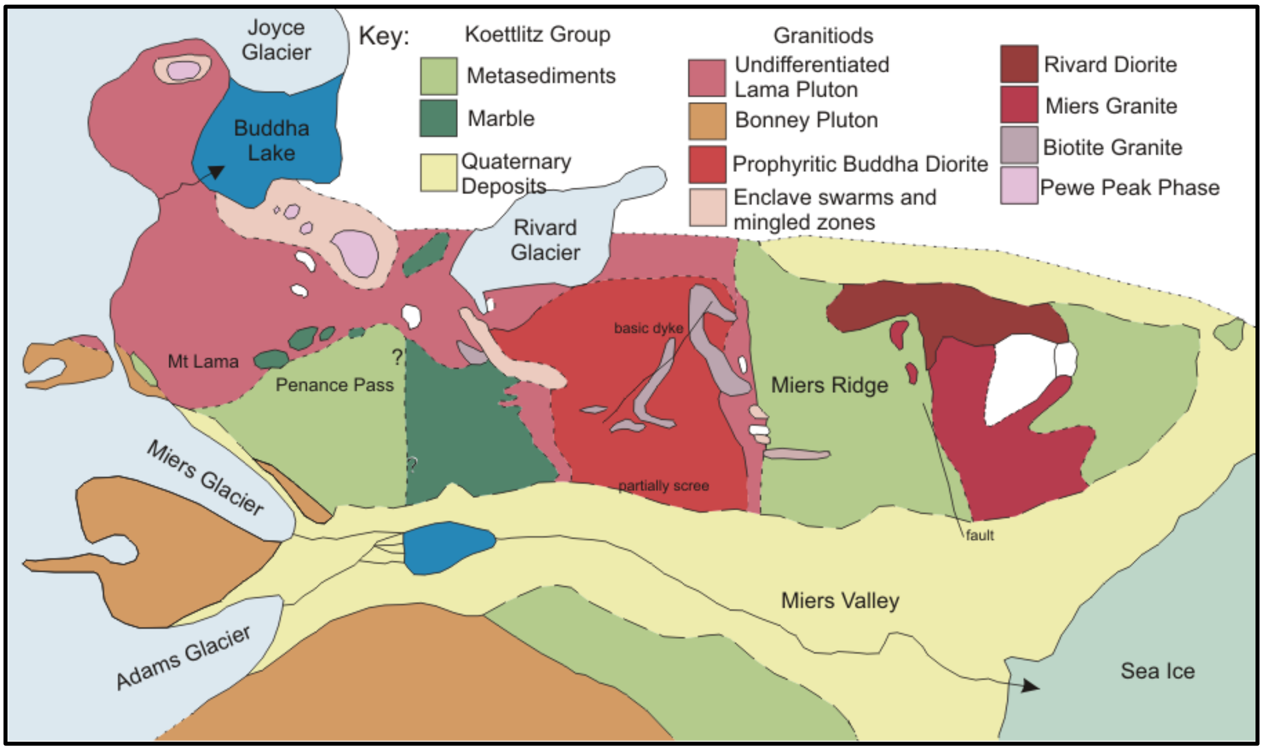

Figure 2.3: A sketch map modified from Worley (1992) showing 8 different granitoids around the Miers Valley.

- Geology of Northern Walcott Bay – Jones (1996)

Completed a study of geology within Walcott Bay (the valley next to the Miers Valley) investigating both the orientation of Quaternary dikes and faults. The study came to the conclusion that most Quaternary dikes have a strike orientation of NE-SW indicating the reactivation of pre-existing weaken zone for initial uplift of the Royal Society Range (Behrendt & Cooper 1991, 1994). As part of the study aerial photographs of the surrounding area where investigated in which a Quaternary fault was found within the Garwood Valley. The dating of the activation of the fault was suggested by the fault offsetting strandlines, similar to ones previous dated by Denton & Hughes (1981). The strandlines were suggested to have formed between 17,000 and 21,200 yr B.P. which suggests the fault must be younger than the strandlines. Jones (1996) indicated by measurements from aerial photographs, the faulting within the Garwood Valley offset strandlines by a sinistral movement of at least 15 m. This fault is later discussed in Chapter 4 where the aerial photograph and a figure are shown (Figure 4.5).

2.3. Geomorphological

2.3.1 Glacial Geomorphology

Early geomorphological studies were conducted by a party led by Péwé in 1957-58 during the IGY. Their studies were directed at investigating the climate variations during the Quaternary period. Péwé published several papers on the field work completed during the 1957-58 field season. This papers included; glacial geology, fluctuations recognised between 1911 and 1958, polygonal ground, multiple glaciations and aging the moraines (Péwé 1958a, b, 1959, 1960, 1962, 1966).

The studies suggested that there are two forms of glacial systems occurring in the area. These were identified as;

1. Drainage glaciers; these are typically large outlet glaciers that originate from ice sheets and drain through the Transantarctic Mountains. They can split from one large glacier into smaller outlets through valleys (e.g. Blue Glacier splits into the Adams, Miers and Joyce which are seen in the field area).

2. Alpine glaciers; these are typically smaller which are restricted to mountain catchments and are independent from the ice sheets.

The oldest glacial deposited that were recognised by Péwé, (1966) were considered to middle Quaternary and only found on high ridges above 600 m.

The glacial history of McMurdo Sound was suggested in 1968 by Denton & Armstrong, where they studied along the coastline of McMurdo Sound and into Taylor Valley. A series of drift sheets are observed around the mouths of the valleys, and terminating usually a small distance into the valleys. This evidence suggested that ice had flowed down McMurdo Sound from the WAIS and had pushed up into the valleys at least four separate times. Within the Denton Hills area these drifts have been labelled M1, M2, M3 and M4 from oldest to youngest respectively. The drifts were named the 'Ross Sea Drift' and had the appearance of a "bathtub ring-line" around the coast of McMurdo Sound (Denton & Armstrong, 1968), indicating the limit of the ice shown in Figure 2.4. The glacial deposits within the Taylor Valley allowed a conclusion that when ice was pushed into the valleys from the WAIS, the EAIS glaciers were smaller. This created the hypothesis that independent glaciers fluctuated 'out-of-phase' of the WAIS.

Figure 2.4: The 'Ross Sea Drift' is observed as contrast black 'bathtub like ring' around the edge of McMurdo Sound at mouths of the valleys. R – Royal Society Range, K – Koettlitz Glacier, B – Adams Glacier, M - Miers Glacier, H – Joyce Glacier, G- Garwood Glacier, D – Rivard Glacier. Péwé (1960), (Photograph by the U.S. Navy, December 6, 1956 Identification No. 00090 F-31 VX6 USN 4/14).

A further study into the fluctuations of ice within the McMurdo Sound region by Denton et al. (1970) identified three major glacial systems operating; WAIS (Ross Sea Ice), EAIS (Flowing from the EAIS sheet through the Transantarctic Mountains) and Alpine Glaciers (Localised mountain glaciers). Each one of these glacial systems form different geomorphic features and fluctuate in different patterns.

In 1981 Denton & Hughes wrote a book on ‘The Last Great Ice Sheets’, including the first dating of geomorphic features. The object of the dating was to constrain a chronology for the advance of the WAIS through McMurdo Sound and pushing into the Dry Valleys. At this time there were limited techniques for dating geomorphology features, limiting it to radiocarbon dating. Due to the lack of carbon sources in Antarctica, proglacial lake deposits had to be used (e.g. dried algae) from lacustrine strandline and deltas. Radiocarbon dating within Antarctica is complicated by the marine water is depleted in ^{14}C due to the upwelling of deep circumpolar water. The marine water is not contaminated by nuclear bomb ^{14}C depleting it by about 16 % compared to global seawater (Stuvier, 1976). This depletion creates anomalous old radiocarbon dates for any material with a marine source.

Denton et al. (1983) further extended their glacial studies, investigating evidence for late Tertiary over-riding of the Transantarctic Mountains. The evidence showed that there were two major imprints of glacial erosion within the Dry Valley region. The youngest event over-rid the Transantarctic Mountains with an ice elevation of about 3000 m, and flowing in a northeast direction. The features that indicate that ice has over-ridden the area are; boulder trains, high mountains with stoss-and-lee shapes, roche mountonnees, fields of subglacial features ripple-like forms and striations. The evidence suggested that glacial over-ridding occurred between 9 - 15 Ma.

Studies using cosmogenic nuclide exposure dating techniques (Brown et al., 1991; Brook et al., 1993) further expanded the hypothesis of ‘out-of-phase’ movement of the glaciers terminating upon land. The studies used detailed mapping of the moraines and the dating of samples within the Arena Valley adjacent to the Taylor Glacier. Brook et al., (1995) used cosmogenic dating for the first time in the Denton Hills field area, sampling the 'Ross Sea Drifts' to further refine an age for the advance and formation of the proglacial lake.

A detailed investigation into the sequence of glacial drifts within the Taylor Valley was conducted by Marchant et al., (1994) to imply glacial fluctuations. The study investigated the cross cutting relationships, outcrop patterns, surface morphology and soil development, to give a general chronology. The detailed mapping allowed four distinctive drifts to be recognised, and by using ^{3}He and ^{10}Be surface exposure dates from Brook et al., (1993), an age for the formation of the drifts. The exposure dating allows a numerical chronology for the sequence of formation of the drifts and the implied ice limits associated with these events. They compared their chronology for the Taylor Glacier with that of other outlet glaciers (Beardmore, Hatherton, Darwin and Reedy Glaciers), they concluded that the fluctuations were minor in the Quaternary and were 'out-of-phase'. The other glaciers showed evidence for expanding during glacial periods throughout the late Quaternary, but the Taylor Glacier showed no evidence for expanding. They concluded that the Taylor Glacier was expanding during interglacial periods throughout the Quaternary and possibly expanding presently.

2.3.2. Lake Deposits

The first recognition of large lakes having previously occupied the valleys was by Péwé (1960) from the preserved geomorphic features and lake sediments. Péwé (1960) suggested that the lakes must have formed when the eastern end of the valley was blocked by ice and/or moraines. Meltwater would have been trapped within the valleys, collecting and forming a lake. During the presence of a lake, deposits such as deltas, shorelines and stratified lake deposits formed. Péwé (1960) identified the remains of a lake within the Miers Valley and called it 'Glacial Lake Trowbridge' and suggested it formed when ice from an expanded Koettlitz Glacier block the eastern end of the valley.

For quite a period of time, the unusual deposits located within the limits of these proglacial lake deposits, along the axis of several Dry Valleys including the Miers Valley, confused researchers. These deposits are mounds and ridges of generally Ross Sea Volcanics material but are further up the valleys than glacial limit shown in Figure 2.5. With the lithology and location of these deposits, researchers purposed that they were somehow deposited by lake processes (Péwé 1960).

Figure 2.5: An unusual deposit found within the pro-glacial lake (Pro-glacial Lake Trowbridge) limit. This deposit features dark 'Ross Island Volcanics' upon light cream/white lacustrine deposits. This deposit is later called a 'cup and saucer' deposits by Clayton-Greene, (1986); Clayton-Greene et al., (1988), Hendy, (2000a, b) and Hall et al., (2006). Photograph from Hendy et al., (2000).

Bradley & Palmer (1967) were the first to purpose possible mechanisms for the formation of these deposits, suggesting two hypothesises. Both mechanisms suggested that the material was transported from Ross Island on WAIS ice across to the valleys, and depositing at the front of the tongues which enter into the valley. During winter the material would freeze into the base of the frozen lake. The sediment is then either lifted to the surface by; compressive forces as the lake freezes faulting and moving the sediment to the surface or alternative theory is ablation causes diapirism which moves sediment upwards. Once the sediment is upon the lake-ice surface it can concentrate creating a mound or ridge, which is rafted during summer further down the lake before melting through and deposited on the lake floor. There was one feature about the mounds and ridges which thee hypothesis by Bradley & Palmer (1967) could not explain. This feature was, that are lacustrine sediments (typical of lake floors) present under the mounds and ridges is but no lacustrine sediment is found within or on the mounds (Figure 2.5).

Further investigation into these features was conducted by Clayton-Greene, (1986) and Clayton-Greene et al., (1988), to determine the origin and processes that had created the features. Clayton-Greene, (1986) and Clayton-Greene et al., (1988) suggested an opposing hypothesis, called a 'lake-ice conveyor' (Figure 2.6), which transported sediment along the top of the lake on the ice away from the terminus of the glacier to the opposite end.

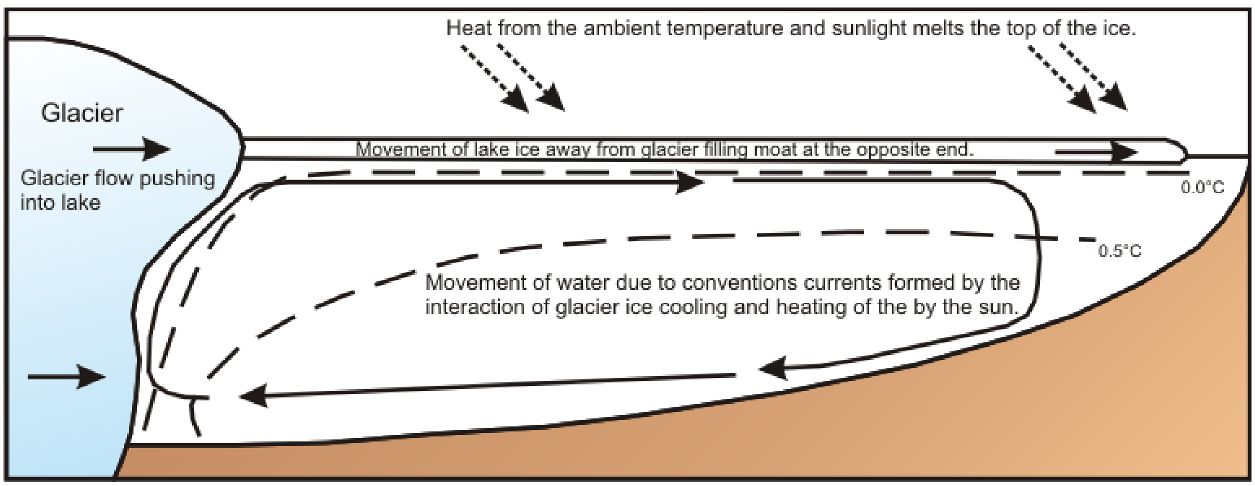

Figure 2.6: Diagram showing the movement of the lake ice due to convention currents, the force of the glacier and the opening up of a moat due to summer melting. Modified from Hendy et al., (2000a).

This theory had been discarded previously as it was thought the sediment would melt/sink through the ice and get deposited. Clayton-Greene (1986) and Clayton-Greene et al., (1988) tested if the ice could transport the sediment by coring Lake Miers, the cores of ice showed that sediment on the ice was restricted to the top 3 - 4 cm. From this it was suggested that the sediment actually protected the ice underneath from the melting during the summer and therefore could not melt through the ice cover on the lake. The 'lake-ice conveyor' mechanism operates by the supraglacially and englacially sediment being deposited directly onto the lake-ice. The movement of the sediment away from the glacier terminus is controlled by the melting of a moat around the lake during the summer. When the moat is formed the lake-ice cover is like a raft with the sediment upon it, and the force of the glacier pushes ice away from the glacier and the moat at the opposite end allows it to move as shown in Figure 2.6.

Over repetitive seasons the sediment is transported on the ice conveyor away from the glacier and into the valley. Eventually the sediment would reach the moat at the far end and get deposited. Sediment is also deposited in small amounts around the sides of the lake creating shoreline ridges, with greater definition than expected in a lake of its size. The mounds and ridges throughout out the valley would be deposited as ice from WAIS retreated, lowering the lake level and depositing the remaining sediment on the floor of the lake. As the mounds are deposited onto the base of the lake they are place on top of the lacustrine sediments, this creates the 'cup-and-saucer' deposit (Figure 2.5).

After Clayton-Greene et al., (1988) published the hypothesis of the 'lake-ice conveyor' mechanism, it gained popularity being considered the most plausible mechanism by the scientific community. This theory was further strengthen by several studies Hendy et al. (2000a, b) studying a modern example of the mechanism and also investigating the interaction of sediment sitting upon the top of lake-ice. Hendy et al. (2000a) used Trough Lake (adjacent to the Koettlitz Glacier), as a modern day equivalent, along with Lyon (1979) study to test the theory. Trough Lake provided a good comparison lake to test the two opposing hypothesis, as it has similar characteristics to 'Glacial Lake Trowbridge'. Trough Lake proved to be too deep (about 90 m) to freeze entirely to base, this means that Bradley & Palmer's (1967) theory of basal freezing could not occur in Trough Lake. In Lyon's (1979) study, the location of a large erratic boulder on a pedestal of ice was mapped and the boulder photographed (Figure 2.7a, b). Hendy (2000a) located the same boulder and measured the distance the boulder had moved in the 17 years between each study. This suggested the boulder had moved on average 18 m per year away for the glacier. This was convincing proof that the boulder could be carried on top of the ice and due to the pressures of the glacier over time, sediment moves away from the glacier. They also observed that on the surface of the lake-ice there were debris and iceberg bands radiating away from the front of the glacier.

The study on the interaction of sediment upon lake-ice showed that material deposited on top of floating ice cannot melt through. The study found that there is a critical thickness is about 5 cm, the isolating properties of any rock greater than 5 cm will protect the ice underneath as shown in Figure 2.8b (Hendy et al., 2000b). If sediment is less the 5 cm the material will melt into the ice and either stay buoyant at depth or melt through the ice cover. Therefore the lake-ice acts like a filter for the 'lake-ice conveyor' mechanism, transporting larger material to the far end and depositing fine material onto the lake floor.

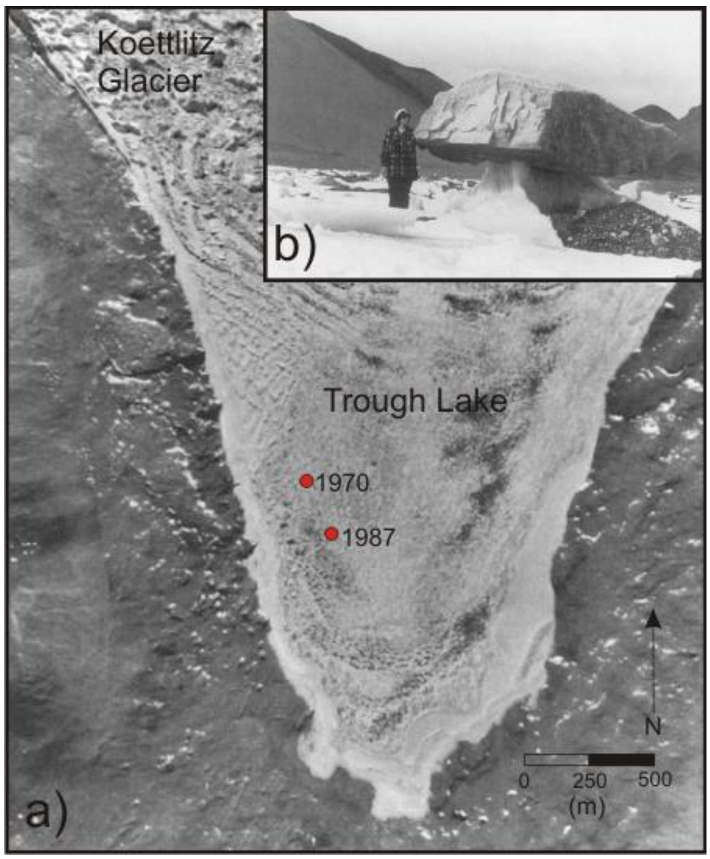

Figure 2.7: a) An aerial photograph of Trough Lake showing the location of an erratic boulder sitting on a pedestal of ice. The boulder was located in 1970 by Lyon, (1979) and then again located by Hendy et al., (2000) in 1987. b) The erratic boulder on an ice pedestal on Trough Lake (Hendy, 2000b). Diagrams modified for original photos.

Dating of the lacustrine sediments have been completed by Stuvier et al., (1981); Dagel, (1984); Judd, (1986); Clayton-Greene et al., (1988); Hall & Denton, (1995, 2000), which allows a chronology of lake development to be created. Hendy (2000a) used this data and new data to investigate the history of lakes within the McMurdo Sound region during late Quaternary. All three (Garwood, Marshall and Miers) valleys were discussed for both the current lakes and pre-historical lakes;

- Within the Garwood Valley, algae was collected and dated from up-thrusted lacustrine sediment by the Joyce Glacier (Stuvier et al., 1978, 1981). The dating of the sample yielded Holocene ages indicating an occupation by a lake after the recession of ice from the Ross Sea, but before the expansion of the Joyce and Garwood Glaciers to their current position.

- No lake currently occupies the Marshall Valley, but the stream draining the Rivard Glacier has exposed sections of lacustrine sediments along its length. These deposits are shown in Figure 2.8, as a sequence of lacustrine and glacial tills.

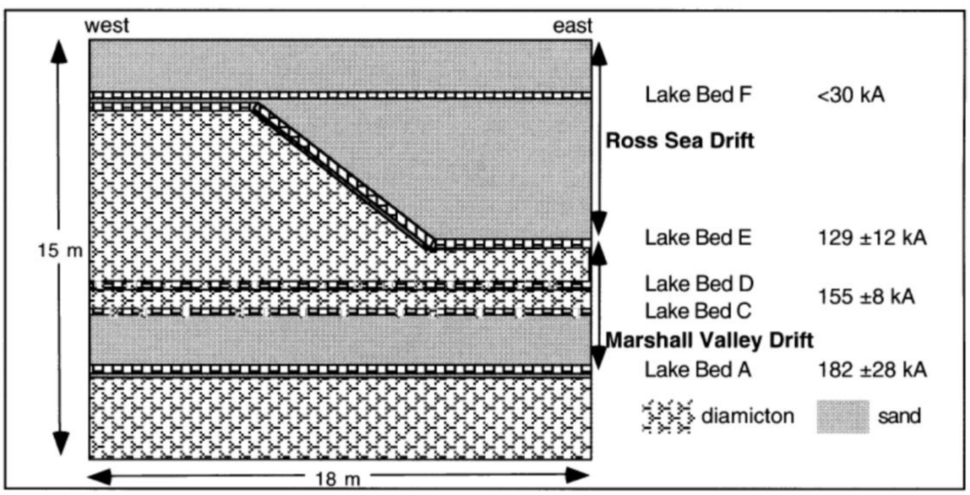

Figure 2.8: A generalised stratigraphy of lacustrine, drift and glacial tills within the Marshall Valley, with average dates for lacustrine beds (Hendy 2000a).

Lacustrine deposits could only be formed by blocking the mouth of the valley by ice from the Ross Sea, forming a lake against the terminus and into the valley. The ages of the lacustrine deposits are therefore directly related to the expansion of ice in the Ross Sea fed by ice coming from the WAIS. The dating allowed the conclusion that ice from McMurdo Sound occupied the valley on two separate periods (isotope stage 2 and 6). During the last occupation lake bed F was deposited (<30 ka) and the previous occupation lake bed A, C, D, E were deposited (between 129 ± 12 and 182 ± 28 ka) shown in Table 2.1.

- As suggested by Péwé (1960) within the Miers Valley there has been a much larger lake than the current Lake Miers. This pre-historic lake has been given the 'Glacial Lake Trowbridge' and was formed by the damming of the Miers Valley. Carbonate and rare algae protected by 'cup and saucer' (Clayton-Greene et al., 1987) features have allowed radiocarbon dating to be used. These dates show occupation of a lake by at least 23,000 ^{14}C yr BP with the maximum extent occurring between 18,000 and 19,000 yr BP.

A collaboration of all dating methods (^{14}C, $^{230}Th/^{234}U$), a summary of changes within the lakes can be placed into a chronology order (Table 2.1).

Year BP	**Method**	**Area**	**Change observed**	**Evidence**
10,000-14,000	^{14}C, $^{230}Th/^{234}U$	Marshall Valley, Lake Miers	Reached levels above present, followed by evaporation.	Carbonate and gypsum evaporates on valley floor (Stuvier et al., 1981; Dagel, 1984; Clayton-Greene et al., 1988).
16,000-18,000	^{14}C, $^{230}Th/^{234}U$	Miers Valley	Lake level up to 450 m above present.	Algae in perched delta, algal limestone on valley floor (Stuvier et al., 1981; Clayton-Greene et al., 1988; Hall & Denton, 1995, 2000)
26,000	^{14}C	Miers Valley	Ross Sea ice sheet meltwater first penetrate valley.	Algae in perched delta, algal limestone on valley floor (Clayton-Greene et al., 1988; Hall & Denton, 2000)
130,000 ± 5000	$^{230}Th/^{234}U$	Marshall Valley	Proglacial lake waters evaporate.	Evaporites on valley floor and beneath „Ross Sea Drift"
145,000 ± 10,000	$^{230}Th/^{234}U$	Marshall Valley	Meltwater from an ice sheet in McMurdo Sound form large pro-glacial lake.	Algal limestone buried in till (Dagel, 1984; Judd, 1986)
180,000 ± 10,000	$^{230}Th/^{234}U$	Marshall Valley	Meltwater from an ice sheet in McMurdo Sound form large pro-glacial lake and then evaporates	Extensive algal limestone/evaporates buried by drift (Dagel, 1984; Judd. 1986)

Table 2.1 Modified from Hendy (2000) showing a summary of changes observed in lakes within in the valleys.

Changes in the presence of lakes and the volume of current lakes have changed dramatically over the Late Quaternary. Hendy (2000) came to the conclusion the lakes throughout the McMurdo Dry Valley region show two different expansion trends relating to the source of water. Small lakes have occupied enclosed drainage basins during high sea level period (isotope stages 1, 5, 7, 9), suggested to be because increased precipitation due to higher temperatures (Hendy, 2000). During the low sea level periods (isotope stages 2, 6 and possibly 8) large lakes have formed in the mouths of many of the McMurdo Dry Valleys. This is suggested that with low sea level the grounding line of the Ross Ice Shelf extends northwards and blocking the drainage systems of the valleys, forming lakes (Hendy 2000).

Once the importance of the 'lake-ice conveyor' mechanism had been recognised, Hall et al., (2006) completed a detailed study summarising the characteristics of 'lake-ice conveyor' deposits. Hall et al., (2006) suggested that most 'lake-ice conveyor' deposits can be characterised by the following:

- Well sorted sand and/or silts, either massive or stratified.
- Reverse grading, with the surface covered by poorly sorted, coarse debris not derived from deflation of the underlying unit.
- Abundance of 0.3 - 1 m high mounds, either in sheets or as isolated features.
- Lacustrine algae.
- Intact pre-existing surface commonly exposed in patches.

In addition, these features also can be present:

- Buried lake ice.
- Well-developed moat line ridges that resemble shorelines.
- Cross-valley ridges, some of which can resemble moraines.
- Sinuous longitudinal ridges.
- Grounding-line mounds and moraine bank.
- Biological precipitates and evaporites.

The 'lake-ice conveyor' is an important mechanism to understand for interrupting Antarctic geomorphology as it explains how drifts and erratics can be deposited further into the valleys than glacier ice reached. It also furthers the understanding into the former glacial limits, water depth and palaeoclimate conditions (Hall et al., 2006).

2.3.3. Alluvial Deposits

Several studies into alluvial processes have been conducted on current river and stream systems. These studies have investigated the quantity of water flow and the potential sediment load the water can move. Alluvial processes have the ability to erode, transport and deposit, depending on the quantity and velocity of the water. McConchie & Hawke (2002) measured the discharge of meltwater through the streams in the western end of the Miers Valley. The streams come directly from the Adams and Miers Glaciers running down well defined streams into Lake Miers. Data of discharge rates were calculated by putting hydrometric control structures in the streams from Adams and Miers Glaciers and also from the outlet stream from Lake Miers. Discharge rates vary throughout the day, relating to the quantity of solar radiation and air temperature shown in Figure 2.9.

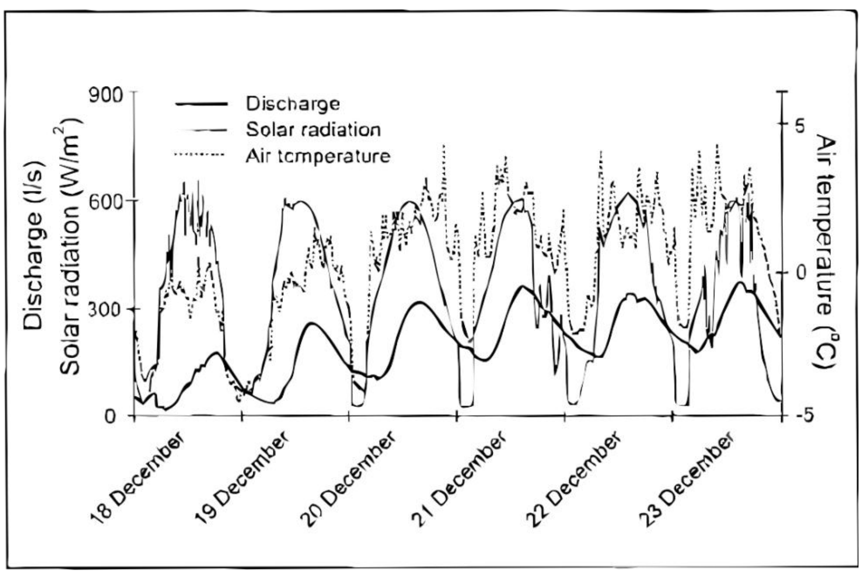

Figure 2.9: Taken from McConchie & Hawke, (2002) showing the effects of solar radiation and ambient air temperature on the discharge during a week. The measurements were taken using 120° v-notch weir on the Adams stream, about 200 m from the snout of the Adams Glacier.

The mean discharge for the Adams and Miers Streams are; 120 and 108 Ls^{-1} respectively and had a maximum discharge of 901 and 820 Ls^{-1} during the period summer of 1989/90 (McConchie & Hawke, 2002). Discharge through the streams changes throughout the day reducing by about half during the 'night' hours, and increasing to a maximum around 1400 NZST (Hawke & McConchie, 2001). The particle size measured from the bedload collected in the Adams and Miers streams, remained the same even with higher discharge rates Figure 2.10. Alluvial processes are an important mechanism for the transport of sediment through the valleys; however it occurs for a limited time during the summer period and is strongly related to amount of solar radiation received by the glaciers.

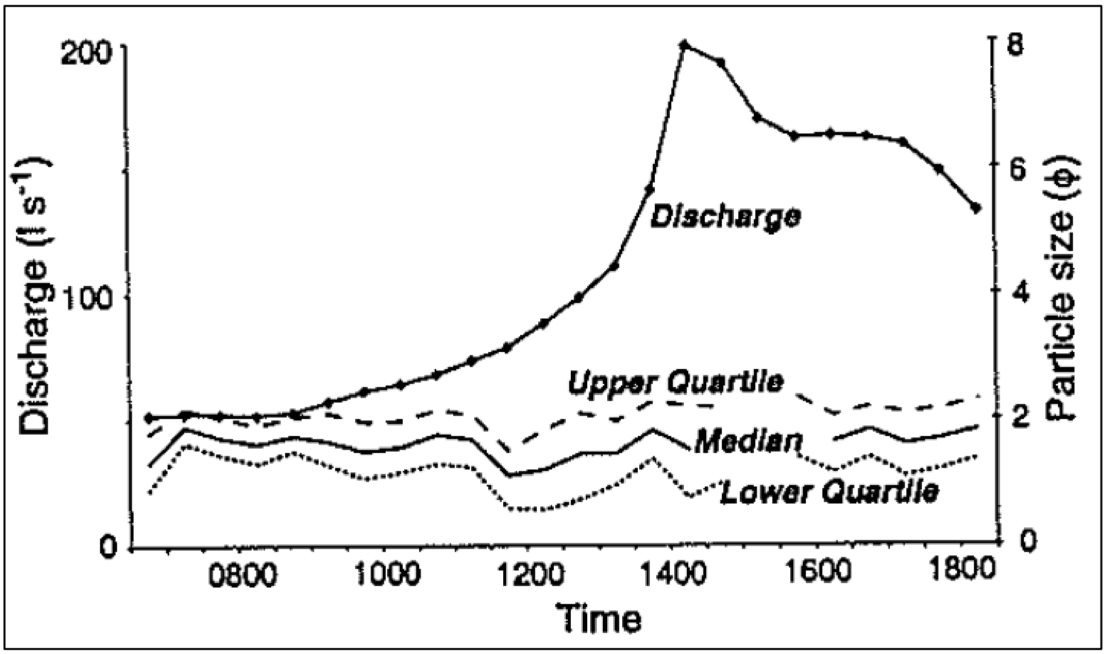

Figure 2.10: Taken from Hawke & McConchie, (2001) showing the particle size of the bedload remains relatively constant during the period of a day, even with increased discharge.

2.3.4. Aeolian Deposits

The extreme winds that funnel through the McMurdo Dry Valley region are an important mechanism for both erosion and deposition of landforms. Recognition of these landforms has been commented upon by Morris et al., (1972) when using the dry valley environment as an analogue for possible Martian landscapes. Studies into the movement of aeolian deposits have been conducted in Victoria, Wright and Taylor Valleys. Victoria Valley had large dunes fields of about 3.5 km long and less than 1 km wide and also flat aeolian sand sheets (Calkin & Rutford, 1974). The movement of the slip faces on the dunes within the Victoria Valley was measured to be about 48 mm/day (Calkin & Rutford, 1974). This dune field is comprised of sand and the measured were during 29 days during summer.

Stronger winds are observed during the austral winter and the early summer as katabatic winds. These winds flow from the EAIS down through the valleys to the coast at wind speeds exceeding 130 km/hr (Nylen et al., 2004). Lancaster, (2004) did a comprehensive study across several valleys to determine the aeolian sediment transport. The study calculated for sand (~250 μm) to be transported 3 m above the surface in Victoria Valley, a wind speed of 9.26 ms^{-1} had to be reached (Lancaster, 2004). This is a relatively low wind speed, which allows a lot sediment to move around and out of the valleys, this is also allows organic to be transported through the valley. The transport of sediment can be increased by 70 % in very cold dry air than in hot dessert environments (McKenna Neuman, 2004).

Meteorological influences controlling the migration of dunes was further discussed by Speirs et al., 2008, investigating the influences of temperature and wind direction. The dune field within the Victoria Valley was the focus of the study, extending the previous studies. Speirs et al., 2008 suggested the threshold wind speed to be 5.3 ms^{-1} ± 0.21 ms^{-1}, which is significant lower than the value calculated by Lancaster, 2004. During the summer months the moisture within the dunes evaporates with the increasing sunlight hours, temperatures of dunes have been recorded as high as +24.5° C (Speirs et al., 2008). As the surfaces of the dunes reach relatively high temperatures, it dries the sediment and removes the moisture. This increases the sediment load of the air as weight of the sediment is reduced and the moisture content of the air is reduced. Speirs et al., 2008 found that the migration of the dunes was far higher than Calkin & Rutford (1974), recording 330 mm day^{-1} on the slip face of the dune.

2.4. Cosmogenic (Surface Exposure Dating)

Cosmogenic dating also known as SED (Surface Exposure Dating) and Cosmogenic Nuclide Dating, is a technique of dating surface features. The technique uses samples collected from the surface, where they have been exposed to the cosmogenic rays produced from galactic sources and rain down through the atmosphere. Cosmogenic rays interact with elements within the crystal structure of the minerals, changing the element into different isotopes. The dating process determines the concentration of these isotopes and with a known production rate, an age can be determined (Discussed in further detail in Chapter 5).

The Arena Valley within the McMurdo Dry Valleys was the focus of Brown et al. (1991), who used ^{10}Be and ^{26}Al isotopes to date moraines. This study was one of the pioneering studies using cosmogenic nuclide exposure dating in Antarctica. The McMurdo Dry Valleys provide a favourable environment for using in situ cosmogenic dating as they are a frozen desert it limits the soil development and chemical erosion (Brown et al., 1991). In the study area most moraines are comprised of sandstones and siltstones derived from the Beacon Supergroup (Brown et al., 1991). The samples were collected from the top of six different moraine crests which represent different fluctuations of the Taylor Glacier (an outlet glacier of the EAIS). The study attempted to constrain a chronology of the fluctuations of the Taylor Glacier. The data provide varied results, the younger moraines were not well constrained yet the older moraines had well constrained ages. This study did test the technique of cosmogenic nuclide dating as a method for producing glacial chronology in Antarctica, proving that it is a feasible method as long the sample are assumed to be in a closed system. Closed systems are when the sample has only undergone a simple exposure history, allowing a simple relationship between the concentration of nuclide and the exposure age.

Brook et al., (1993) further extended the use of cosmogenic nuclide dating in the McMurdo Dry Valleys by using the Brown et al., (1991) data as subset and adding addition ^{10}Be data and ^{3}He isotopes. This study placed defined ages for four different moraines within the Arena Valley. The study used the two isotopes ^{10}Be and ^{3}He to compare against each other to constrain the chronology of the formation of the moraines. The ^{3}He dates for the moraines are; 113 ± 45 ka, 208 ± 67 ka, 335 ± 187 ka and 1200 ± 200 ka for the Taylor II, III, IVa and IVb moraines respectively (Brook et al., 1993). The ^{10}Be ages produced corresponding ages for the Taylor II and III but showed older ages for Taylor IVa and IVb. Moraines represent the glacial limit for which the Taylor Glacier had expanded into the Arena Valley. Brook et al., (1993) data supported ‘out-of-phase’ behaviour of the Taylor Glacier, expanding during interglacial periods (Figure 2.11).

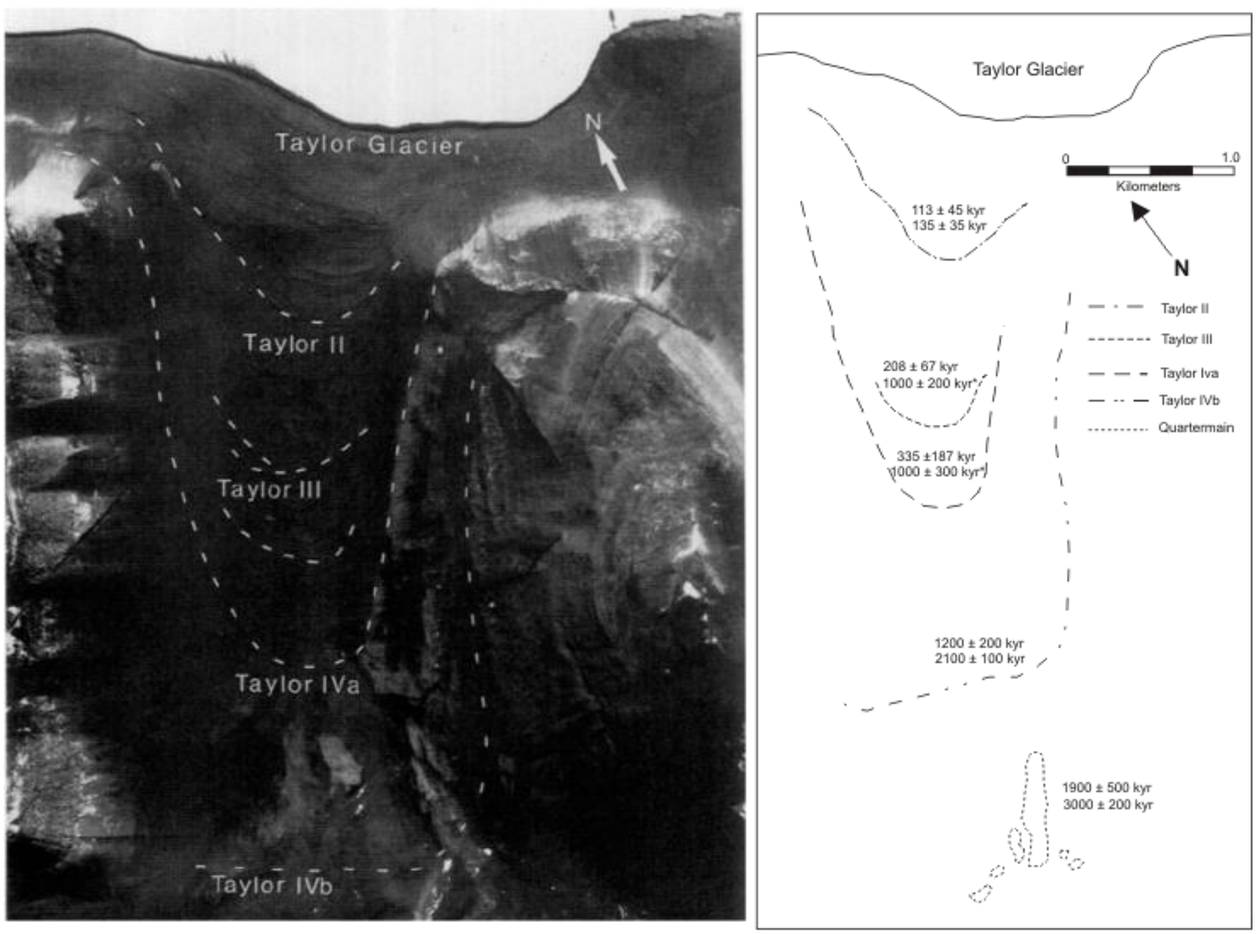

Figure 2.11: Diagram of the moraines and drifts within the Arena Valley. The ages shown on the figure are ^{3}He (top) and ^{10}Be (bottom) from Brown et al. (1991) and Brook et al. (1993). A general pattern of the youngest closest to glacier and oldest further away is observed, although the two isotopes show different ages for the older material. Figure modified from Brook et al. (1993).

The first use of surface exposure dating within the Denton Hills study area was completed by Brook et al., (1995). The focus of this investigation was to provide another method of dating the Ross Sea Drift, as previously the drift had been dating by radiocarbon dating of algae fossilised from the proglacial lakes. The Ross Sea Drift is predominantly comprised of dark volcanic material is well persevered around the coast of McMurdo Sound. A combination of ^{3}He, ^{10}Be and ^{26}Al cosmogenic nuclide was used to date the material and then compared to the radiocarbon dates. The cosmogenic ages found from the moraine material were similar to the radiocarbon ages of the fossilised algae. The carbon ages from Denton & Hughes (1981), Judd (1986), Denton et al., (1989), Dagel et al., (1990), suggest that there was a lake occupying the Miers Valley between about 10,000 and 24,000 ^{14}C yr B.P. Brook et al. (1995) cosmogenic data suggest the moraine was formed between 8,000 and 33,000 years, from sample location within the Miers Valley and Blue Glacier areas. This study tested the technique of using cosmogenic nuclide dating as a valid form of determining Antarctic glacial chronology, and with the similarity between the ages of radiocarbon dating and cosmogenic dating it was proven to be a valid method.

3. Geology

3.1. Introduction

Basement geology throughout the McMurdo Dry Valleys has been studied since the early scientists on the expeditions of Scott and Shackleton. Further studies have extended the knowledge into the regional extend of the units (Gunn &Warren, 1962; Blank et al., 1963) and into the specific lithologies (Allibone, 1988; Cox, 1989; Smillie, 1989; Worley, 1992). Presented here are the summary lithologies from field observations of the mapped geological units on Appendix Sheet 1.

3.2. Methodology

The surface geology was mapped within the field area by generalising the specific lithologies into the main units. This allowed a greater area to be covered in the time allowed, and without chemical and/or microscopic techniques some of the lithologies are indistinguishable. The mapping was conducted over a two week field season. The geological boundaries, units and structures were drawn directly on base maps/images.

After the ground mapping of the main units was completed, other small scaled features were added by using aerial photographs of the valleys. These aerial photographs provided an insight into the areas that were inaccessible and/or features missed by the ground field surveying, for example small scale dikes which crop out throughout the area.

3.3. Koettlitz Group

The Koettlitz Group has been identified as the oldest formation throughout the McMurdo Dry Valley area and comprises of a variety of metasediments (Blank et al., 1963). These metasediments are suggested to have formed from originally marine sediments which were deformed by a series of events destroying many of the original sedimentary structures (Allibone, 1988). This regional metamorphism deformed the sediments into the amphibole facies (amphiboles are dominate in many of the lithologies), which indicates medium to high grade regional metamorphism occurred. The metasediments have undergone at least three phases of deformation (Allibone, 1988). Garnets within the metasediments continued to grow after the regional deformation ceased, indicating a slow cooling and uplift (Blank et al., 1963).

For this thesis the Koettlitz Group has being separated into three units (gneiss, schist and marble) each with a general description below.

3.3.1. Gneiss

Gneiss was dominantly found in the northern and southern edges of the field area, on the southern valley side of the Miers Valley and on the northern wall of the Garwood Valley. The gneiss is thought to be the oldest lithology due to the metamorphism stage and grades into schist. The gneiss appears as the massive unit in the valley sides, outcropping as vast areas. Adjacent to and often folded into the gneiss is the Salmon Marble.

In some areas the visual appearance of gneiss is very similar to schist, to separate the two lithologies; the gneiss has been defined as having compositional layering. The gneiss displays none of the original sedimentary structures. Compositional layering forms within rocks due to high grade metamorphic processes. These processes allow a semi-ductile separation of minerals into mafic and fewer felsic layers, shown in Figure 3.1. Often micas (biotite or muscovite) form the boundary between mafic and felsic minerals.

In hand specimen the gneiss appears as a layered dark grey rock, with fewer thin cream/white layers. It is a predominantly a medium-coarse grained groundmass with occasional large felsic porphyroblasts ranging up to 20 mm in diameter. The dark layers contain mainly mafic minerals; biotite, amphiboles, garnet and the felsic layers contain plagioclase and minor quartz. Felsic layers are often less than 10 mm thickness and usually non-continuous, pinching out as they are traced along. The porphyroblasts of plagioclase have undergone semi-ductile metamorphism deforming the porphyroblasts into sigma clasts.

Figure 3.1: This gneiss block shows compositional layering, most of the block is comprised of the darker mafic minerals with the lighter feldspar bands at the top right corner. Larger white/cream porphyroblasts of feldspars are seen contrasting against the darker groundmass.

Weathering on the gneiss occurs at different rates depending on the compositional layer, the more resistant minerals such as the feldspar withstand erosion and protrude out of the fine grained groundmass. In the block above (Figure 3.1) the light cream/white feldspar are seen protruding from the surface. The gneiss will discolour to a reddish/brown colour from the oxidation of minerals within the mafic layers; this discolouration can be seen on the base of the gneiss block in Figure 3.1.

3.3.2. Schist

Schist within the Denton Hills area is limited to a few small areas; high on the ridge between the Miers and Marshall Valleys, and a small amount in the north-western end of the Miers Valley. The schist has a very similar appearance and mineral assemblage to the gneiss, therefore has been defined as having a well-defined cleavage/schistosity. Cleavage/schistosity is observed as thin laminations of near uniform thickness throughout the rock, often formed by the quantity of mica along the boundaries. The schist often shows a later phase of deformation, being folded after the formation of schistosity.

The appearance of schist is typically a dark grey, with differing hues due to the variation in mineral assemblages and weathering. An example of the differing hues is observed in samples with a large quantity of biotite, in a fresh sample it displays a purplish hue and when weathered displays a reddish brown hue.

Schists are fine grained with only the largest minerals (biotite) reaching up to 3 mm in size. The fine grainsize makes the appearance of the schist darker. Additional minerals are; garnet, quartz, amphibole (e.g. hornblende, tremolite). Within in the schist there are an interbeds of different mineral assemblage schists and rare thin interbeds of marble.

The schist weathers to a brown-grey colour from the oxidation of the biotite. The weathering also emphasizes the layers of the schistosity, moisture collects along the schist planes and freezes pushing them apart and breaking the schist apart.

3.3.3. Marble

The marble throughout the area is probably the most recognisable unit, seen outcropping as a cream/pinkish white contrasting against the darker units (Figure 3.2). It is second most abundant lithology outcropping throughout the area, often as folded and twisted bands. In fresh outcrops the marble is seen as a brilliant white, some outcrops have greyish colour due to impurities, finer crystal size and known graphite layers (< 5 mm). Due to these impurities, when weathered the marble can turn to a cream or pinkish colour. This pinkish colour found in some areas, resulted in the unit being called 'Salmon Marble' and was the reason behind the naming of Salmon Bay just north of the Garwood Valley.

Figure 3.2: In the front of the photograph is the brilliant white marble, contrasting against the reddish/brown granitoids in the distance. Near the ridgeline of the southern wall of the Garwood Valley.

When observed from a distance, large scale bedding marked by the grey impure layers, can be seen folding and twisting into complex forms. This bedding has been suggested to show the original sedimentary bedding (Blank et al., 1963) and shows the intensity of the deformation the area has undergone.

In hand specimen the marble is generally coarsely crystalline, with individual crystals ranging in size between 2 - 4 mm. Faint small scale bedding structures can be seen often marked by graphite or aligned mica layers. The fewer darker grey/cream layers are due to a fine crystal size and/or impurities such as the graphite and micas.

When the marble is weathered it often turns pinkish/red from impurities within the marble such as iron rich minerals (e.g. biotite) which are oxidised to iron oxide (rust). Staining from weathered granites can also alter the appearance of the marble. The surface of weathered marble is relatively smooth, as the crystalline marble weathers uniformly.

3.4. Granitoids

Granitoids are the most dominant lithology within the study area, dominating the western (Royal Society Range) side and the ridges around the Marshall Valley. Most of the granitoids throughout the McMurdo Dry Valleys are known under the Supergroup of the 'Granite Harbour Intrusives' complex. Within this Supergroup is a large variety of lithologies, due to; style of emplacement, mineral assemblages and timing of emplacement. Many of these different granitoids have been identified by either chemical and/or microscope analysis (Smillie, 1989; Worley, 1992).

Granitoids have been mapped as one unit, 'Granite Harbour Intrusives', including all the differing lithologies (Appendix Sheet 1). There are a range of emplacement styles throughout the area, ranging from large elongated plutons to small (< 0.5 m) dikes (Appendix Sheet 1).

In hand specimen the differences between some of the granitoids are obvious differing in; colours, crystal sizes and mineral compositions. Some of the differences are because of the different phases and styles of emplacement. The colour ranges from bluish/grey through to greyish/red, but the most common is whitish/grey which oxidises and weathers to reddish/brown.

The crystal sizes differ immensely within the granitoids ranging from coarsely crystalline granites through to fine crystalline diorites. The reason for the differing crystal sizes is mainly due to the method of emplacement; smaller fine crystals are in thin dikes and larger coarse grained crystals are in larger pluton like emplacements. Granitoids appear to cut across the metasediments, suggesting a younger age of emplacement. The absence of 'baked zones around the granitoids indicate that the granitoids were emplaced while the metasediments were still relatively hot.

Different granitoids have differing fabrics, some showing preferred orientation of crystals within the unit. This indicates partial crystallisation of the melt before it flowed into the emplacement, the flow would orientated and crystals long the flow direction. The fabric can be used to help interrupt the emplacement history. Other larger emplacements are homogeneous displaying no fabric, usually representing the larger plutons where the melt has been emplaced before crystallising.

The granitoids have been severely weathered in the high ridgeline areas, creating astonishing tafoni as shown in Figure 3.3. Most of the granitoids are also weathered to the reddish/brown colour and tends to create a coarse surface to touch, as the weathering has emphasized the coarse crystals.

Figure 3.3: Granitoids showing the extreme weathering on the ridgelines. The reddish brown colour is due to oxidising of iron within the minerals.

3.5. Mafic Intrusions

Mafic intrusions have been emplaced throughout the area after the formation of the other lithologies. The mafics are basaltic in composition, being very fine crystalline dark black/grey colour with the rare larger crystal of olivine. The largest emplacements of these mafics are seen around Penance Pass – Shangri-La area, as several larger bands.

More commonly the mafics are seen as numerous basaltic dikes scattered throughout the area, often in stark contrast against the lighter colour units of marble and granitoids (Figure 3.4). The basaltic dikes swarms cross-cut all of the other bedrock geology indicating that they are the youngest unit. The age of the dikes can be estimated as within the Ross Sea region there have been many Quaternary basaltic deposits and emplacements similar to these identified. When mapped the dikes throughout the area, trend NE-SW (Appendix Sheet 1), which correlate to the study of Quaternary dikes and faults in Walcott Bay by Jones (1996).

Figure 3.4: A mass of dikes on the northern wall of the Garwood Valley. Here the black basaltic dikes are in stark contrast to the white marble bedrock. (Photo courtesy of nzTABS project).

The dikes are dark grey/black fine grained rock, usually are fresh surfaces with rare weathered surfaces to brownish/black. The dikes range in thickness between < 10 cm up to a few metres and can be seen transecting across valley walls, disappearing underneath the Quaternary deposits. The dikes tend to crop out as straight lines across topography, suggesting that they are sub-vertically dipping.

3.6. Deformation

Throughout the field area the basement geology has been deformed by both folding and faulting. Deformation has folded the metasediments of the Koettlitz Group, creating complex outcrop patterns throughout the area. Faulting is observed offsetting lithologies, for example in the Marshall Valley on the north wall marble has been faulted a number of times creates offsets of the marble. These faults display offsets of a few hundred metres and are at a high angle striking north-northwest.

Figure 3.5: A section of the northern wall at the western end of Lake Miers shows the deformation of the Koettlitz Group, dominantly marble with minor schist inclusions.

Later faulting minor faulting occurred in area such as the ridge between the Marshall and Garwood Valleys. These faults appear as high angle near vertical faults which offset lithologies, dikes and geomorphic features. On the northern valley wall of the Marshall Valley the marble has been faulted in several places, offsetting it by only a few meters. On the southern valley wall of the Garwood Valley opposite the Garwood Glacier, there is a series of surface lineations which have been faulted by approximately 15 m (Jones, 1996).

3.7. Quaternary Deposits

Throughout the Denton Hills area the valley floors are filled with unconsolidated deposits, which for the geological map have categorised as unconsolidated units. These deposits are further examined and discussed in the following chapter and the geomorphological map (Appendix Sheet 2).

4. Glacial Geomorphology

4.1. Introduction

The McMurdo Dry Valleys are series of valleys initially created by alluvial processes (Sugden et al., 1999) which drain the Transantarctic Mountains east into McMurdo Sound. These valleys have been modified by glacial processes, creating both erosional and depositional features that have preserved. Discussed below are geomorphological features that were observed in the field and mapped, creating the first detailed geomorphological map for the Denton Hills area.

4.2. Methodology

4.2.1. Field Geomorphic Mapping

Initially geomorphic features were drafted on to base maps by using the ALOS (Advanced Land Observation Satellite) images, LIDAR (Light Detection and Ranging) data and previous maps and studies. These base maps were taken down to Antarctica during the 08/09 season to ground truth. Field mapping was conducted over two weeks in November. This was a short amount of time, limiting the area in which could be covered. The field work was conducted by walking around the area and investigating areas of interest for ground truthing. In the field features were drawn directly onto the base maps and photographs were taken so similar features could be identified. As part of the field work, over 450 photographs were taken from the side of a helicopter of the valleys sides and floor.

4.2.2. Digitising Geomorphic Maps

The final completed map has been produced by the amalgamation of all the sources of information/data. This includes previous studies, photographs, and field observations of other members within our team and interpretation of similar features. Similar depositional deposits have similar colouring/shading and erosional features tend to have symbols/lines. The map has been created by using Arcmap and CorelDraw programmes.

4.3. Lakes, Rivers and Streams

Within the study area, there are three lakes and associated drainage systems of streams and rivers which actively transport the meltwater during the summer months (Late November till Mid February) Hawke & McConchie, 2001. During the field season (Early November) the hydrological system was frozen with very little liquid water. Ice cover over the lakes is multiyear ice with evidence of a moat forming at the edge of the lake during the warmer summer months.

4.3.1. Lakes

Buddha Lake is situated in the west of the study area, between the Miers and Garwood Valleys and behind the Marshall Valley. The Joyce Glacier flows around Péwé Peak and into the northern end of the lake, restricting the outflow of the lake. Meltwater from the Joyce Glacier is the main source coming directly into the lake at the north and via a small stream at the west. A small amount of inflow to the lake is sourced from small ponds and seasonal melt from the surrounding hill slopes. Buddha Lake is the largest lake in the study area measuring 1.6 km by 1 km, occupying 1.6 km^2. The outflow from the lake flows through a small gap in the north-eastern corner of the lake, down into Lake Colleen.

a)

b)

c)

Figure 4.1: a) Photograph of Lake Buddha, towards the north with the Joyce Glacier wrapping around Péwé Peak.

b) Photograph of Lake Colleen, towards the northeast with the Garwood Glacier in middle right.

c) Photograph of the Lake Miers towards the east, taken from Holiday Peak with the Miers Glacier on the left.

In the Garwood Valley, Lake Colleen sits at the western end of the valley, between the Joyce and Garwood Glaciers. The lake is fed by a stream from Buddha Lake, meltwater from the Joyce Glacier, the Garwood Glacier and permanent snow patches on the valley sides. The lake drains east around the terminus of the Garwood Glacier and towards McMurdo Sound. The Lake Colleen is approximately 1 km long and 0.5 km wide, occupying about 0.5 km^2.

Lake Miers is situated at the western end of the Miers Valley and is fed by the meltwater from both the Miers and Adams Glaciers. The outlet of this lake is east down the valley to McMurdo Sound via a relatively large river. The Lake Miers is approximately 1.25 km along and 0.6 km wide, occupying about 0.8 km^2.

4.3.2. Rivers/Streams

Associated with the lakes there are a network of rivers which transport the inflow and outflow of water. There are also a network of smaller streams which transport meltwater from the valley sides and permanent snow patches. For most of the year many of these streams are either bare or frozen, but during the summer months they transport the meltwater around the landscape and deposit it into either lakes or out to McMurdo Sound. River and streams have the ability to deposit, erode and transport sediment through the landscape. Depositional features consist of the river terraces, bedform and supply of sediment to deltas. Erosional features presently active in the landscape are the formation of channels, cutting through older deposits and some bedrock meltwater streams. Sediment is transported through the valleys as bedload in the rivers and streams. Hawke & McConchie (2001) discussed how the bedload of the Miers River (flowing between the Miers Glacier and Lake Miers) was related to the volume of meltwater from the glacier and this was related to the amount of solar radiation. They also mentioned that there were influxes of sediment when the banks of the stream collapsed in and the frozen sediment was released into the flow.

Figure 4.2: A photograph of the meltwater stream coming out north side of the Joyce Glacier into the Garwood Valley. The stream has a series of terraces associated changing flow with the stream.

4.4. Polygonal Patterned Ground

Polygonal patterned ground (PPG) forms as a result of a thermal regime within permafrost. Permafrost is defined as, earth material which has an average annual temperature of less than 0.0°C for more than two consecutive years. As the ground freezes and thaws the ground expands and contracts which form a network of cracks. The cracks form polygonal shaped areas and the cracks fill with sediment stopping them from refreezing over. Over repeated events the cracks deepen and become well defined borders (Godfrey et al., 2008).

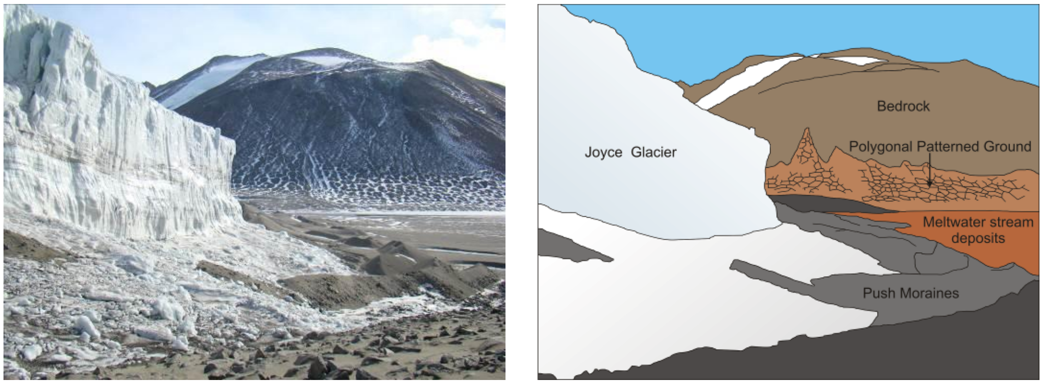

Figure 4.3: a) Photograph showing polygonal patterned ground on the side of the Garwood Valley. The photograph is taken across the front of the Joyce Glacier, in the west end of the Garwood Valley. b) A sketch of the photograph highlighting the polygonal patterned ground.

PPG is found within the study area along the valley floors predominately on drift sheet deposits and on the sides of the valleys on the colluvium deposits. Drift sheets provides a relatively flat and stable surface for PPG process to form over successive seasons. As PPG forms in areas of frozen sediment, PPG has started to from on the colluvium on the side of the valleys. Varying stages of the formation of PPG can be identified on different sedimentary units, which may show the evolution and a chronology for the emplacement of the sediment.

4.5. Desert Pavement

The term desert pavement is used to describe the flat smoothed ground surface that formed by erosion to form one uniform smooth level. The erosion is dominantly wind abrasion, moving sand and ice particles along the surface and removing fine material from the sediment. The desert pavement appears as a compacted coarse grained surface, of interlocking stones (Figure 4.4). The wind erosion removes the fine material leaving an armouring surface of coarse pebbles and cobbles protecting the sediments below.   As the ‘pavement’ is a veneer the reality is that the underlying deposit is unconsolidated and is ‘soft’ under foot. As you walk across the deposit the weight and movement of your feet can break through the surface ‘pavement’ surface and expose the fine material being protected beneath.

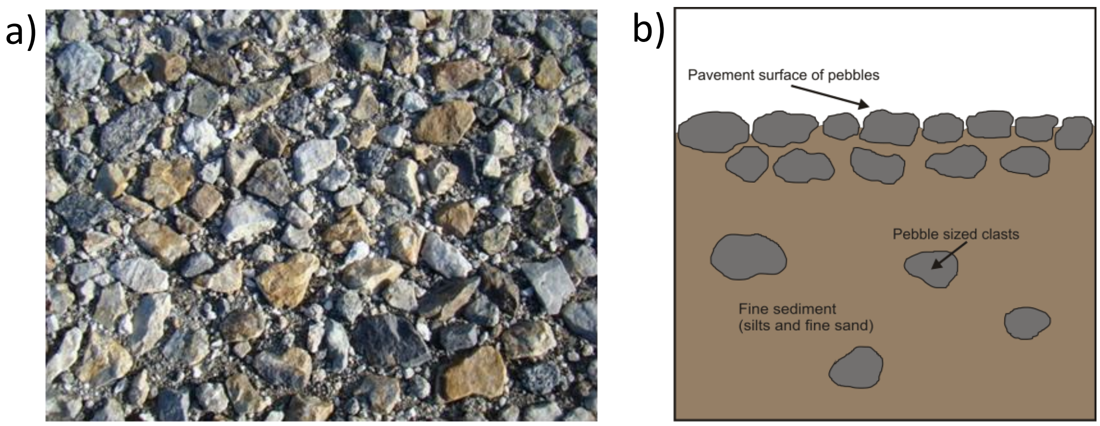

Figure 4.4: a) A photograph of the typical appearance of the desert pavement, which covers most of the area. The top is observed as interlocking pebble to cobble sized clasts, with granules and coarse sand infill the gaps. b) A sketch of a cross section through the top layer of the desert pavement, large clasts at the surface with fines (e.g. silt & sand) protected underneath. There are the rare larger pebble size clasts under the surface.

4.6. Quaternary Faults

For the geomorphology we have only expressed the ‘recent’ faults observed within the study area. Our definition of the term ‘recent’ is faults that are seen within the landscape to offset geomorphic features. Faults have been identified in all of the valleys, offsetting lithological layers however these are much older, as they do not offset younger glacial deposits (these have not been mapped on the geomorphic map). The timing of the faulting has not been well defined, and is only restrained when the faults offset geomorphic features.

Jones (1996) completed a study into the Quaternary faulting within the northern Walcott Bay area, which also included a Quaternary fault identified in the Garwood Valley.   The Quaternary faults through this area, all tend to strike NE-SW and offset dikes and geomorphic features. The

Garwood fault identified on the southern valley side, opposite the Garwood Glacier (Figure 4.5), was said to have displaced lacustrine strandlines, formed by a pre-historic lake. Offset of linear features are obvious in aerial photos however the interpretation of them being lacustrine strandline may be incorrect. The features are at an elevation of about 750 m which is much higher than the post-glacial lake that formed within the Garwood Valley is thought to have reached.

There is evidence for some offset of the ridgelines between the valleys, moving in a sinistral motion. This motion is the same sense as that of the smaller identified fault and the faults within the Walcott Bay area studied by Jones (1996).

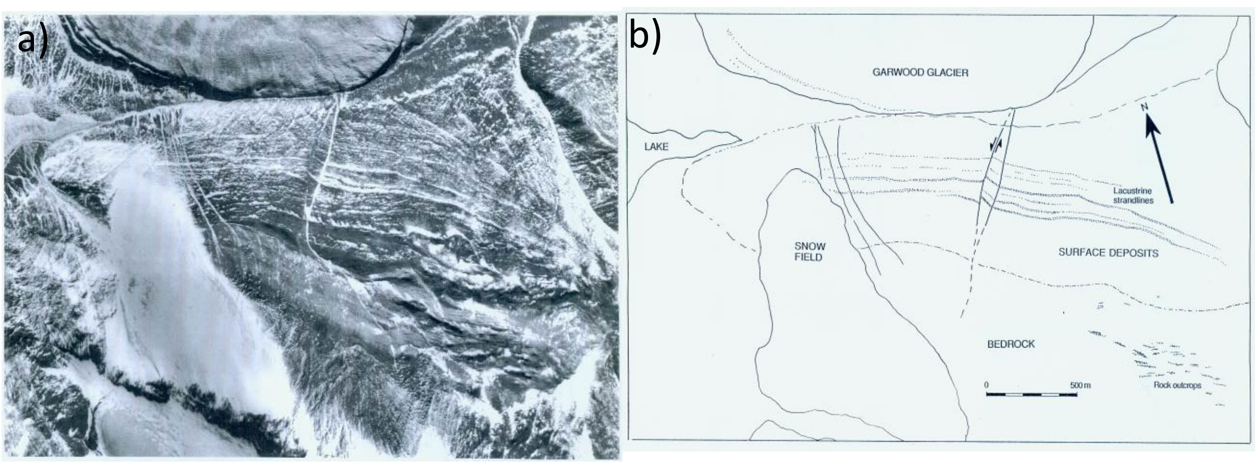

Figure 4.5: a) Aerial photograph taken of the south side of the Garwood Valley, opposite the Garwood Glacier (From Jones, 1996, obtained from the US Navy). b) A sketch map drawn by Jones, 1996 showing the interpretation of a sinistral offset linear surface features.

4.7. Mass Movement

4.7.1. Alluvial Fans

Throughout the study area alluvial fans are present transporting material from the steep slope to base of the valleys. Alluvial fans use the meltwater that is present during the summer months to move sediment down slope, depositing it into a radiating fan shape on to the floor of the valleys.

The alluvial fans have formed over the top of glacial deposits, which clearly shows the relationship of the fans being post-glacial features. Most of the fans are still actively growing and grading into a base level of either a lake or into the valley floor.

In the Garwood Valley, there are a larger number of alluvial fans than in the other valleys. Within the Garwood Valley there are several patches of permanent snow on both sides above Lake Garwood/Colleen, the meltwater produced by these over the summer months would provide enough water to move sediment. Around Lake Colleen the alluvial fans (Figure 4.5) grade into the lake level, in the eastern end of the valley the fans grade into the base level of the valley floor ending as the slope shallows.

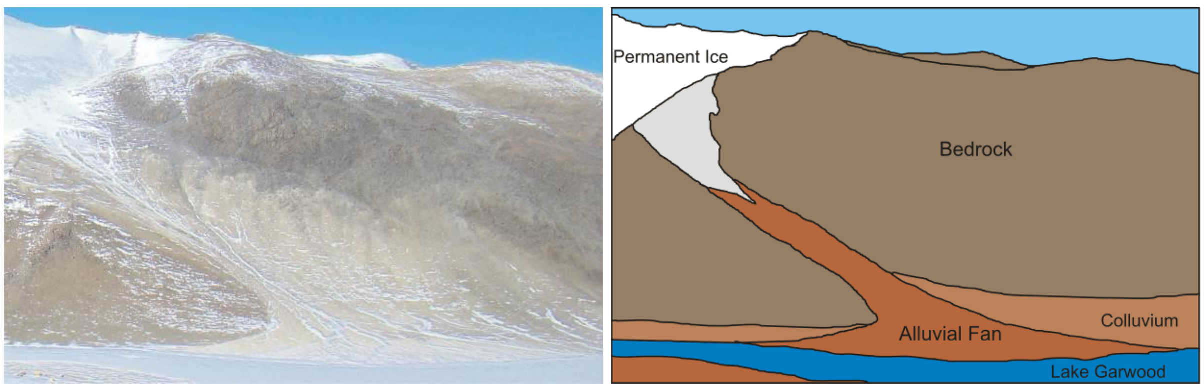

Figure 4.6: a) A photograph of the southern wall of the Garwood Valley. An alluvial fan is fed by the permanent ice field at the top of the photograph. The fan then flows down and terminates into Lake Colleen, the change in colour of the rock is due to a change in lithology. b) A sketch diagram provides a clearer interpretation of the alluvial fan and also shows the colluvium that forms along the side of the valleys.

Within the Miers and Marshall Valleys there are alluvial fans that grade into the valley floors. There are fewer fans and only tend to form when there is a steep slope or large catchment area for water to be supplied from.

4.7.2.   Colluvial Deposits

Colluvial deposits have formed at the base of the valley walls throughout the study area. The colluvium forms by erosion of material higher on the slope, that moves down to the base of the slope by gravity. The colluvium is comprised of angular material with the same lithology as the bedrock directly above the deposit. Colluvium has started to cover the drift sheets, moraine sequences and proglacial lake deposits, which indicates that the colluvium is post-glacial and is an active process presently.

4.8. Wind Processes

4.8.1. Aeolian Deposits

Aeolian deposits are present throughout the McMurdo Dry Valleys, forming dune fields and other deposits. The formation of the deposits is a very active geomorphic process throughout the area, creating and modifying landforms. All studies into aeolian deposits within the McMurdo Dry Valleys suggest it a very dynamic system that reacts to meteorological conditions (Speirs et al. 2008). Wind-speed reaches very high speeds through the narrow valleys, which increases the volume of sediment that can be transport within the airflow. A study by McKenna Neuman (2004) showed how very cold and dry air can also transport up to 70% more sediment than hot dessert environments. The aeolian deposits within the study area are comprised of coarse sand, which is derived from the crystalline rocks which make up the bedrock geology. Crystalline rocks are susceptible to extreme physical erosion, producing a large volume of sand (Hawke & McConchie, 2001). The sand within the study area is deposited in areas where it is protected from getting reworked. The sand covers the topography where it is deposited, covering over the top of geomorphic features.

At the front of the Joyce Glacier there appears to be a dune field, where dunes have formed in the lee of the glacier's terminus. The dunes appear to be 10-15 m high ridges which encircle the front of the Joyce Glacier. However these are push moraines at the front of the Joyce Glacier, which are identified by a cross section created by a drainage river cutting through at the north end. These push moraines are covered by 5-30 cm layer of aeolian sands, covering the underlying landform and creating high ridges. Lower ridges (less than 5 m high) further away from the glacier terminus may indeed be dunes that have formed adjacent to the moraine sequence. Ripple features have formed on the aeolian surface by the wind (Figure 4.7).

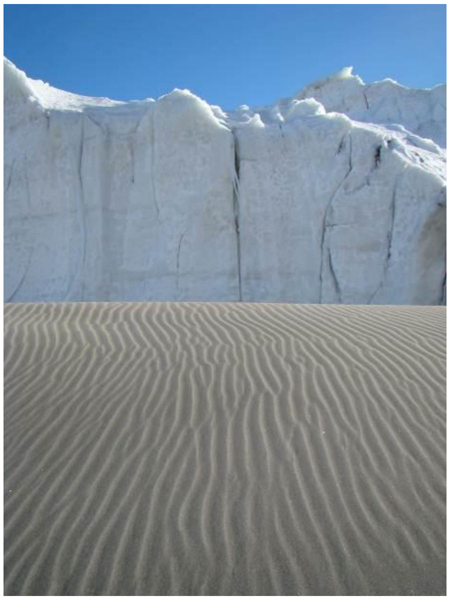

Figure 4.7: Photograph taken of the push moraine ridges at the front of the Joyce Glacier. The moraines are covered in aeolian sands, in this photograph the sand has been formed into ripples by wind blowing across the surface.

There are several large patches of aeolian deposits that have formed in sheltered areas high up on the valley sides. The sediments are usually of sand size material and dominantly quartz-rich grains. Some of the deposits are measuring about 10,000 m^2 and but most of the aeolian sediment is found deposited in between boulders on the slopes of the valleys.

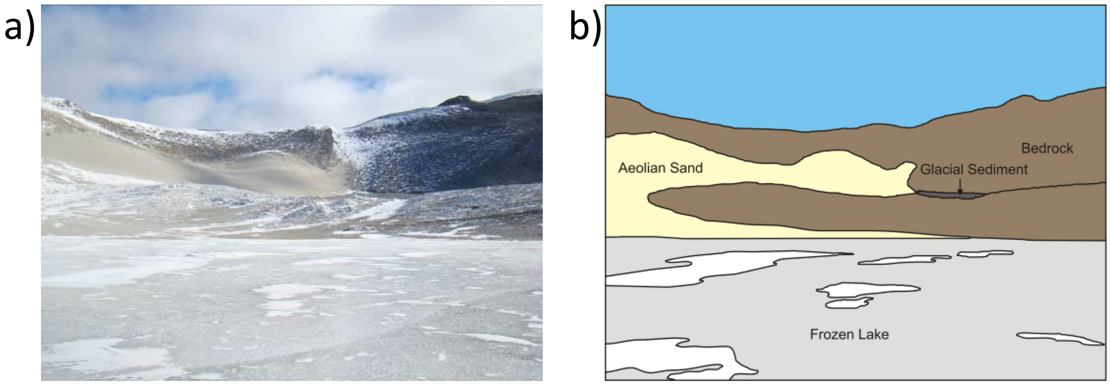

Figure 4.8: a) Photograph taken high up on the ridgeline between the Garwood and Marshall Valleys. The area is a cirque now filled with glacial sediment and a small lake. The white bedrock material is marble and the surrounding darker rock is granite. b) A sketch to indicate the area covered by the aeolian sand.

4.8.2. Wind Erosional Features

High on the ridges surrounding the valleys, the bedrock has been actively eroded by the high winds funnelling through the valleys. The bedrock has been carved into tafoni by the wind action scouring out the rock, and eroding it into precarious shapes. Much of the rock has the characteristic cave-like or pitted, smooth concave holes carved into them. Tafoni is created by sand/small particles being moved around by the wind, once a small pit is created the sand will sit in there and as the wind flow over the boulder the sand will rotate around.

Figure 4.9: a) A photograph taking along the ridgeline around the top of the Marshall Glacier. The granite bedrock has been eroded by wind process to form tafoni. b) A photograph of a boulder in the Miers Valley with the top surface pitted due to the abrasion by windblown material.

There is also evidence for ventifacts, which is the formation of smoothed faces from erosion by sand, dust and ice crystals. There are ventifacts with only one smoothed surface (einanter), which point towards the southwest in the Miers Valley, the dominant wind direction. There are also fewer ventifacts with three sides smoothed into a triangular block (dreikanter).

4.9. Glacial Processes

4.9.1. Depositional

4.9.1.1. Dump Moraines

Dump moraines are created adjacent to the ice by supraglacial material being 'dumped' off the top ice to from ridges. Dump moraines can be divided into two types; terminal, forming at the front of the glacier as it terminates in the valley, and lateral, forming along the valley side adjacent to the ice margin. Dump moraines, size and shape are dependent on the amount of supraglacial material and the standstill period. The moraines tend to be made of unsorted angular material, ranging from boulders (4 m+) to silt.

Both terminal and lateral dump moraines are well preserved around the Garwood Glacier. Lateral moraines have formed one distinctive ridge, along both sides of the valley that the glacier flows from. The terminal moraine has formed into three distinctive moraine ridges on the western side of the glacier and only one ridge on the eastern side. The terminal dump moraines associated with the Garwood Glacier are approximately 40 m high and 100 m wide.

In the Miers Valley, some remnants of dump moraines remain from when the Adams and Miers Glaciers were further advanced. When the remnants of the moraines are mapped, it can be observed that they show a convex shape, crossing over the valley. They also follow the same shape as the present day glaciers terminus.

Figure 4.10: a) A photograph of the lateral dump moraine along the side of the Garwood Glacier. Large boulders are transported on top of the glacier and then get dropped off either off the side or at the terminus.

b) Photograph taken looking north of the terminal moraines at the end of the Garwood Glacier, there are three distinct ridges.

c) Photograph taken looking north along the second moraine ridge out from the Garwood Glacier, SED samples were taken from boulders along the moraines.

4.9.1.2. Push Moraines

Push moraines are formed as the glacier has advanced down the valley, ‘pushing’ sediment in front of the glacier creating mounds. These mounds tend to be comprised of reworked glacial sediments, including outwash sediments. Within push moraines deformed sedimentary bedding can be observed if a cross section through the moraines can be seen.

Around the front of the Joyce Glacier there are push moraines which have deformed the sediment in front leading into Lake Colleen. There are a series of mounds radiating around the front of the glacier, with the largest reaching about 30 m high. At the north end of the mounds, a meltwater stream has cut through and created a cross section through the mounds showing the internal structure (Figure 4.11). The cross section is through mainly coarse sand sediments, however in other area it can be observed that the mounds are comprised of larger material (up to boulder size). In the cross section bedding planes can be observed that are now deformed, forming a surficial non-tectonic reverse (thrust) fault.

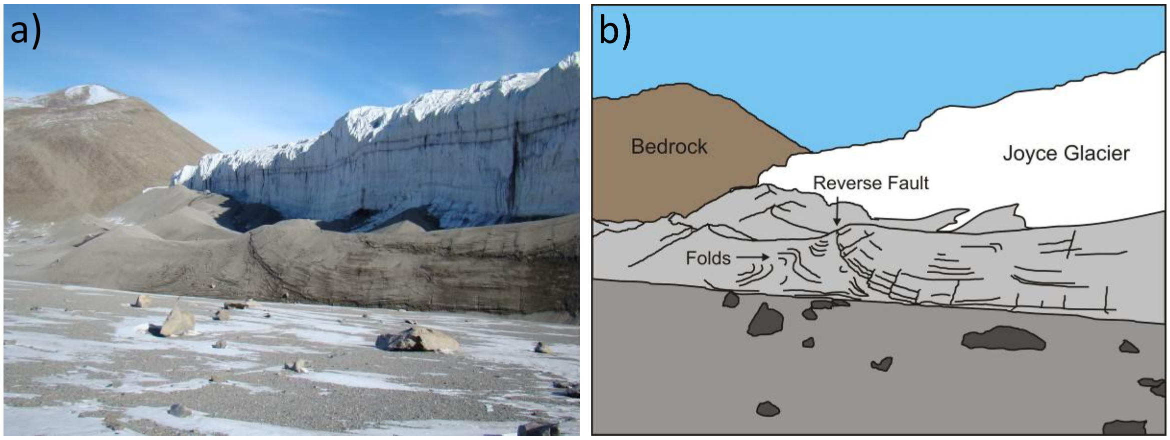

Figure 4.11: a) A photograph of the push moraines at the front of the Joyce Glacier. b) A sketch diagram showing the internal structure of the moraines. Push moraines are created by compression of sediments in front of the glacier creating folds and reverse faults.

4.9.1.3. Drift/Hummocky Moraine

Drift/hummocky moraine is used as the term for glacial sediments left as the glacial has retreated leaving a layer of material in the base of the valleys. Drift sheets can be identified by having different characteristics than the surrounding deposits. These characteristics can be; lithology, angularity, weathering etc. and by looking closely at the deposits, different glacial events can be recognised.

In the east ends of valleys there are deposits of black volcanic material, which has been derived from the Ross Sea Volcanics. Several different glacial events can be identified by weathering and how consolidated the deposits are. The material has been called the Ross Sea Drift (Blank et al., 1963, Péwé 1966) and is found encircling round the mouths of the McMurdo Dry Valleys. Drift sheets are made of sediments of differing sizes and is unconsolidated. The youngest of the drifts sheets is ice-cored which is still melting creating an undulating surface scattered with kettle holes.

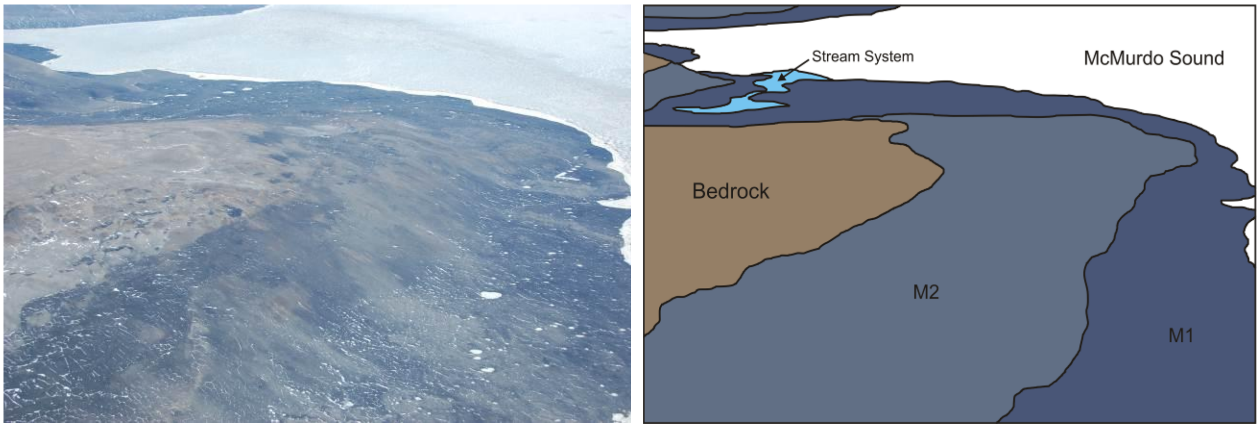

Figure 4.12: a) Photograph taken from a helicopter over the mouth of the Miers Valley, both the M1 and M2 surfaces visible. The M1 surface is the lower darker surface with kettle holes and displays polygonal ground, the M2 is the darker patchy surface rising up the hillslope. The lighter cream/brown is granite bedrock. b) Sketch diagram of the same photo showing the areas covered in M1 and M2 drifts.

The Ross Sea Drift can be separated into two distinctive drift sheets with the area:

- The younger of the sheets (M1), displays an undulating surface of unconsolidated volcanic rich sediments. In some areas of the field area the drift sheet can be observed to be ice cored, which has created a pitted surface with many kettle holes forming. The surface of the drift is dominantly coarse grained but with little disturbance, fine grained material is found under the top layer. From aerial photographs this deposit is very distinctive as a dark black pitted deposit with many ice/melt ponds scattered across the deposit.

- The older of the drift (M2), has a more consolidated surface, and is follows the topography. M2 is comprised of the same volcanic material, but either due to erosion processes or a thinner layer having been deposited, the drift appears patchy. In area the underlying granites and metasediments sediments are visible through the drift sheet.

4.9.1.4. Erratic Boulders

Erratic boulders are boulders that have been carried by the glacier and then stranded when the ice has retreated (Figure 4.16). The boulders are deposited at the limits of the ice, either from the sides or at the front of the glacier. As the boulder define the limit of the ice, they are commonly used to as a marker for glacial limits and used for cosmogenic exposure dating. Identifying erratic boulders, lithology and weathering surfaces provide identifying characteristics as often they differ to the surrounding bedrock.

Within the Miers Valley there is line of erratic boulders half way up the northern valley side. The boulders display a much younger weathering surface and form a line shallowly dipping towards the east. The boulders display relative fresh blue/grey granite surfaces and sit upon the more weathered orange/red granite bedrock lithology. Amongst the erratic boulders there is also one large (about 0.7 m) basaltic boulder, which is different lithology than is observed within the area.

4.9.1.5. Kettle Holes

Kettle holes form usually in hummocky deposits as ice gets incorporated into the sediment. As this ice melts out, the overlying sediment settles into the void space, creating a depression. In the study area the Ross Sea Drifts are scattered with kettle holes, the youngest drift (M1) is still ice-cored many more kettle holes could form. Some of the kettle holes within the youngest drift (M1) have ponded meltwater, these thaw and freeze created ice domes (Figure 4.13).

Figure 4.13: Photograph of a kettle hole within the youngest Ross Sea Drift moraine drift. The kettle hole is filled with meltwater which has frozen creating an ice dome.

4.9.2. Erosional Features

Ice has created many features within the valleys, as ice has helped created the valleys themselves (Denton et al., 1983; Sugden et al., 1999). The original alluvial V-shaped valley was eroded to a glacial U-shaped valley by the processes of glaciation and abrasion. The U-shaped valley has very steep sides and a flat floor, the ice also straighten the valley profile, as ice is more viscous than water. Over the many glaciations the valleys have been carved into straight west-east flowing valleys, draining the Royal Society Range towards McMurdo Sound.

Trimlines are observed on the side of the valleys, at the elevation of the ice as the glacier flowed down the valley. As the ice is forced against the valleys sides, it erodes the side to be forming linear features sloping down valley at the elevation of the ice. The lines dip with the height of the glacier, and terminate when the pressure of the ice on the valley sides, is reduced to where it cannot erode. Ice acts as a good abrasive and can also contain rock fragments which can add to the erosive power. Within the study area there are linear features associated with both the outlet glaciers (east dipping) and from when ice has pushed up into the valleys (west dipping).

4.10. Glaciofluvial Features

4.10.1. Proglacial Lake Benches and Lake Deposits

Within the study area there are proglacial glacial lake deposits, which formed during a period when the eastern end of the valleys were blocked by ice from the Ross Sea. A pre-historic lake was identified in the Miers Valley by Péwé (1960) and was suggested to be named Glacial Lake Trowbridge (Lake limit shown on Appendix Sheet 2). Later studies also indicated that smaller lakes had formed in both the Garwood and Marshall Valleys (Denton & Hughes, 1981). During the occupation of the valleys with the lakes, the lakes created benches/shorelines which now indicate the limit of the lake and the recession series as the lake drained. Many of the lake benches contain remnants of the carbonate and gypsum layers which were formed in the lake.

In the Miers Valley, Glacial Lake Trowbridge formed when the eastern end of the valley was dammed by ice from the Koettlitz Glacier and the advance of ice into the valleys from the Ross Sea/ McMurdo Sound embayment (Péwé, 1966). The lake then formed from the meltwater filling the eastern basin of the Miers Valley, as the climate warmed the quantity of meltwater increasing the size of the lake to fill both the eastern and western basins. A detailed study was conducted by Clayton-Greene (1986) showing the extent of the lake and the associated geomorphology features. Clayton-Greene et al., (1988) suggested that the lake reached at least 156 m above sea level from carbonate deposits. The lowest carbonate deposits have been identified at 77 m above sea level, which suggests the lake at the largest was about 80 m deep (Clayton-Greene et al, 1988).

Within the Garwood Valley there is less preserved evidence for a large proglacial lake occupying the valley for a period of time. There is evidence for a small lake forming in the eastern end of the valley. There is a well preserved delta complex in the eastern end, which has been dissected by the present river. The delta has a series of levels stepping down from the highest elevation of 43 m above sea level, to presently a surface about 24 m above sea level. Preserved within a cross section of the delta are sedimentary structures of topsets, foresets, and bottomsets (Figure 4.14).

Figure 4.14: a) Photograph of a river cut cross section through the delta in the Garwood Valley. The cross section allows the internal structure to be observed. b) A sketch showing the internal structure of the delta.

The Marshall Valley has very limited evidence for a relatively small lake forming in the eastern end of the valley. The Marshall Valley is the smallest valley in the study area and has a steep gradient from McMurdo Sound. There are a few outcrops/deposits of carbonates and gypsum in lower reaches of the valley which indicated that for a short period of time a body of water occupied the area.

4.10.2. Lake Ice Conveyor Deposits

Clayton-Greene (1986) suggested a mechanism to form unusual deposits in the base of the valleys (Figure 4.15). The deposits are; sinuous ridges at the west end of the valleys and mounds scatted randomly along the valley. The deposits are comprised of the same material as the drift sheets in the eastern end of the valley, black volcanic basalts, but are a lot further up the valleys than the evidence suggests the ice advanced to. Clayton-Greene (1986) showed how the ice on the top of a proglacial lake could transport glacial derived material further up into the valley than the terminus of the glacier. As the ice terminated in the lake, any supraglacial debris would fall upon the ice cover of the lake, creating a ridge of material on the lake ice. During the summer melt period a moat would form around the lake and due to the force of the glacier, the lake ice would move away from the glacier. Over the summer period, the deposits would move to the west end of the lake and get deposited as a ridge. The mounds found in the valleys have been described to be material that had been on top of the lake ice, which was lowered to the floor as the lake drained.

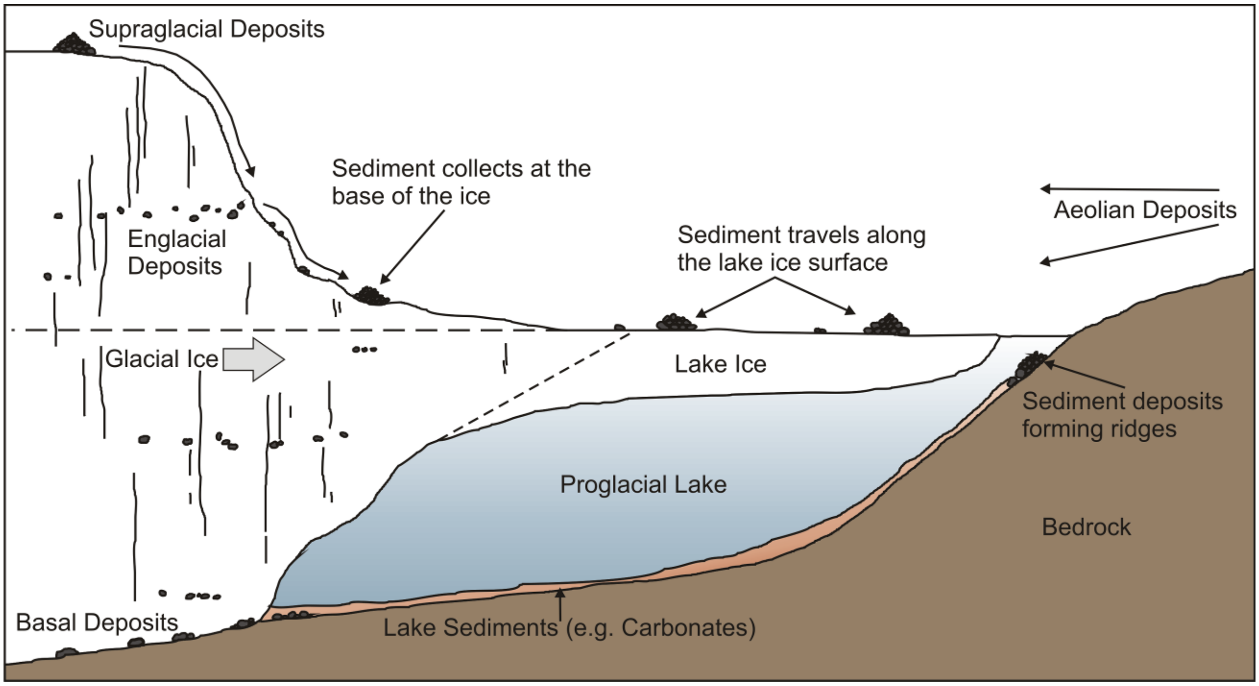

Figure 4.15: This diagram of the lake-ice conveyor presents the possible sources and methods of transporting glacial derived sediment further up valley than the glacial limit. Supraglacial and englacial sediment is deposited at the glacial ice/lake ice margin. During the summer the ice at the edges of the lake melts forming a moat, the pressure from the glacier pushes the lake ice away towards the other end of the lake. Over several annual cycles the sediment reaches the moat at the far end and gets deposited as a ridgeline into the lake (Modified from Clayton-Greene, 1986).

The hummocky deposit of drift material between the two basins within the Miers Valleys is typical of a lake ice conveyor deposit. The proglacial lake would have incorporated the two basins until this hummocky deposit was laid down, separating the western basin from the eastern basin and from the additional water.

When the mounds were lower to the base of the lake, another deposit was formed. The deposit was named ‘cup and saucer’ deposit, and was a mound of the black volcanic drifted material sitting upon a white carbonate or gypsum rich lake sediment layer. These deposits are directly linked to being a lake deposit as they incorporate the lake sediments.

Preserved in the field area at the western end of the Miers Valley are a series of low topography ridges (Figure 4.16). These ridges are characteristic features of the lake-ice conveyor mechanism, which often have the appearance of moraine ridges. They form either as simple curve or straight or sinuous ridges which trend cross the valley.

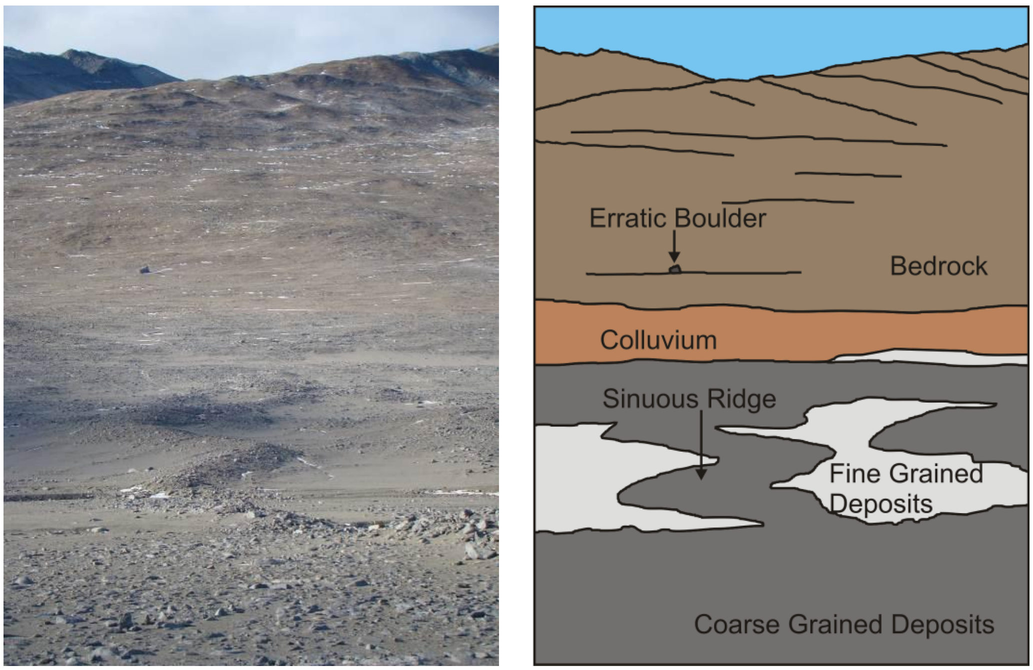

Figure 4.16: a) A photograph of a sinuous ridge at the western end of the Miers Valleys looking across from the southern side. b) Sketch of the photograph, indicating the sinuous ridge across the valley floor. Also observed in the sketch is an erratic boulder halfway up the hill-slope and also trimlines in the bedrock.

5. Terrestrial in-situ Cosmogenic Nuclide Dating

Introduction

Terrestrial in-situ cosmogenic nuclide dating is a technique which allows age constraints to be assigned for the period of time a rock surface has been exposed at the surface. These age constraints are determined by the concentration of naturally occurring isotopes formed by the interaction between the rock surface and cosmic rays from space. As this technique dates the exposure period at the surface, it is also known as surface exposure dating (SED). The methodology is explained in further detail below.

Surface exposure dating is widely used to date geomorphological features and is commonly used for dating glacial material associated with both glaciation and deglaciation (Gosse & Phillip, 2001; Cerling & Craig, 1994). Surface exposure dating was used in this study to date several geomorphic features throughout the Denton Hills area, allowing time constrains to be placed on the formation of glacial features complimenting other geomorphic evidence into the reconstruction of glacial fluctuations.

5.1. Background of Cosmogenic Nuclide Dating

5.1.1. History

In-situ cosmogenic nuclide dating is a relatively new dating technique, originally being proposed as a theoretically possible dating method in 1955 by Davis & Schaffer. Davis & Schaffer (1955 & 1956) proposed the technique by discovering that ^{36}Cl (Chlorine-36) isotope concentrations increased within rocks with altitude. This allowed the theoretical proposal of using cosmogenic nuclide as a method of dating exposure ages however the concentrations of the cosmogenic nuclides needed to be measured were exceedingly small, far beyond the limits of instruments of the time.

The development of highly sensitive conventional mass spectrometry and accelerator mass spectrometry (AMS) in the 1980's allowed isotopes ratios as low as 10^{-15} to be measured. This technical advance allowed the first in-situ cosmogenic nuclide dating to be measured. This advance was followed by a rapid expansion of the technique, as much of the theoretical background (e.g. production rates, shielding, geomagnetic controls) had been studied by physicists. The technique soon expanded from the initial Chlorine (Cl) isotopes, which had been discovered by Davis & Schaffer (1955) to include isotopes of Helium (He), Beryllium (Be), Carbon (C), Neon (Ne), Aluminium (Al), Chlorine (Cl), Manganese (Mn) and Xeon (Xe).

The first use of using in-situ ^{10}Be exposure dating was accidental, when a research team from the University of Pennsylvania came across a large component of in-situ produced ^{10}Be in basalt flows along the Columbia River (Brown et al., 1982). They continued to study ^{10}Be empirically and determine production rates of ^{10}Be and ^{26}Al within quartz in a combined study with Nishizumi et al., (1986). The terrestrial ^{10}Be exposure dating was first applied to date the exposure of Libyan Desert Glass from Egypt (Klein et al., 1986).

With the use of the technique expanding, in-situ cosmogenic nuclide dating surged into science literature in 1986 with six papers covering the rare-gas isotopes (Craig & Poreda, 1986; Klein et al., 1986; Kurz, 1986; Lal, 1986; Nishizumi et al., 1986; Phillips et al., 1986). The technique has been applied to many geomorphological studies, as geomorphic landforms are often difficult to date, as they are made from reworked material.

This technique has been critical in defining glacial chronology on both global and regional scales as other dating techniques are often restricted by the physical characteristics of glacial deposits. Previously, most glacial events were dated by using radiocarbon dating, requiring the presence of carbon based material. Glacial deposits rarely have carbon based material associated directly with the glacial event. Often the sources of carbon (e.g. logs, branches or algae) used to date glacial events would have been reworked. SED allows the direct dating of boulders or an eroded rock surface which directly related to glacial events.

Within Antarctica, SED has already been used by several researchers across the continent, providing new insight into the behaviour of ice fluctuations. Two previous studies have used cosmogenic dating within the McMurdo Dry Valleys to assist in defining the timing of glacial events. Although both projects studied a much larger area, a few samples were tested from the Denton Hills area. Brown et al., (1991) used ^{10}Be and ^{26}Al to study moraines within the Arena Valley, to determine the timing of glaciation to compare it to global glacial cycles. The other study was by Book et al., (1995) who used ^{3}He to study moraines throughout the area, including some samples from the Denton Hills area. The samples tested from the Denton Hills area were chosen to define the movement of ice from the WAIS, represented by the Ross Sea Drifts, in the east of the valleys.

5.1.2. Theory

The principle behind cosmogenic nuclide dating is that the cosmogenic isotopes are produced at a known rate and also decay at a known rate. Cosmogenic isotopes are created when a rock exposed at the surface interacts with cosmogenic rays, this interaction creates the isotope and the concentration of altered isotopes increases with the period of exposure. By measuring the concentration of these cosmogenic isotopes, and applying the known production rate and decay rate an estimation of how long the sample has been exposed can be determined.

Cosmogenic rays consist of high energy particles (neutrons, protons and subatomic particles such as muons). These rays are from either solar (Sun) or galactic (outside our solar system) sources, which constantly bombard the earth. As these rays come through the earth's atmosphere they interact with the particles within the atmosphere, reducing the concentration. This reduces the concentration of cosmogenic rays and consequently the number of interactions with rock surfaces with lower altitudes. On reaching the rock surface, cosmogenic rays only have the energy to penetrate a few centimetres into the surface. On penetration into the rock if they interact with an atom they can cause a 'spallation' (refer to Figure 5.1) reaction changing the make-up of the atom and creating the cosmogenic isotope.

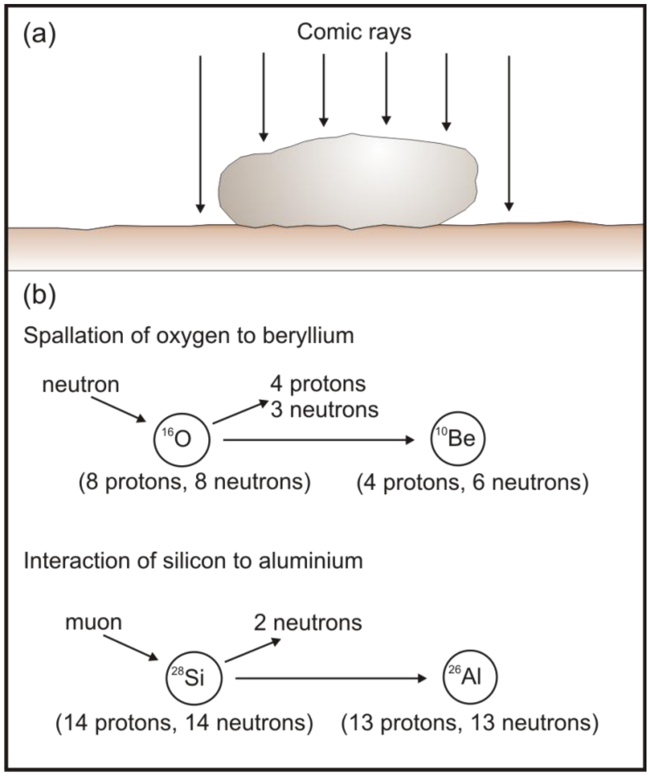

Figure 5.1: (a) Cosmic rays consisting of neutrons, protons and subatomic particles (e.g. muons). (b) The interaction between the cosmic rays and quartz. Oxygen reacts with a neutron to produce an isotope of beryllium; silicon reacts with a muon to produce an isotope of aluminium (Modified from Manz, 2002)

A 'spallation' reaction is when a high energy particle from the cosmogenic rays collides with the nucleus of an atom within the mineral lattice of a rock, changing the atom into a cosmogenic isotope. The change occurs because the collision changes the atomic mass, by removing both neutrons and protons (e.g. ^{16}O changes to ^{10}Be and ^{28}Si changes to ^{26}Al). These isotopes are radioactive, decaying over a period of time (e.g. ^{10}Be half-life is 1.36×10^6 and ^{26}Al half-life is 7.17×10^5) which needs to be accounted for when estimating the exposure age (Cerling & Craig, 1994).

The formation of cosmogenic isotopes produces very small concentrations. To analyse concentrations in these very small quantities, the instruments need to be very precise. Modern AMS (Accelerated Mass Spectrometer) instruments enable these very small concentrations to be measured (Figure 5.2). AMS works on the principle that charged particles will be deflected along different paths when passing through a magnetic field depending on their mass. The test sample is placed at one end of the instrument where it is ionized and sent through the instrument. As the ionized materials passes through the instrument, it passes through two magnetic fields. The magnetic fields alter the path of the charged particles, sending them to separate analysers which count individual atoms (Hellborg & Skog, 2008).

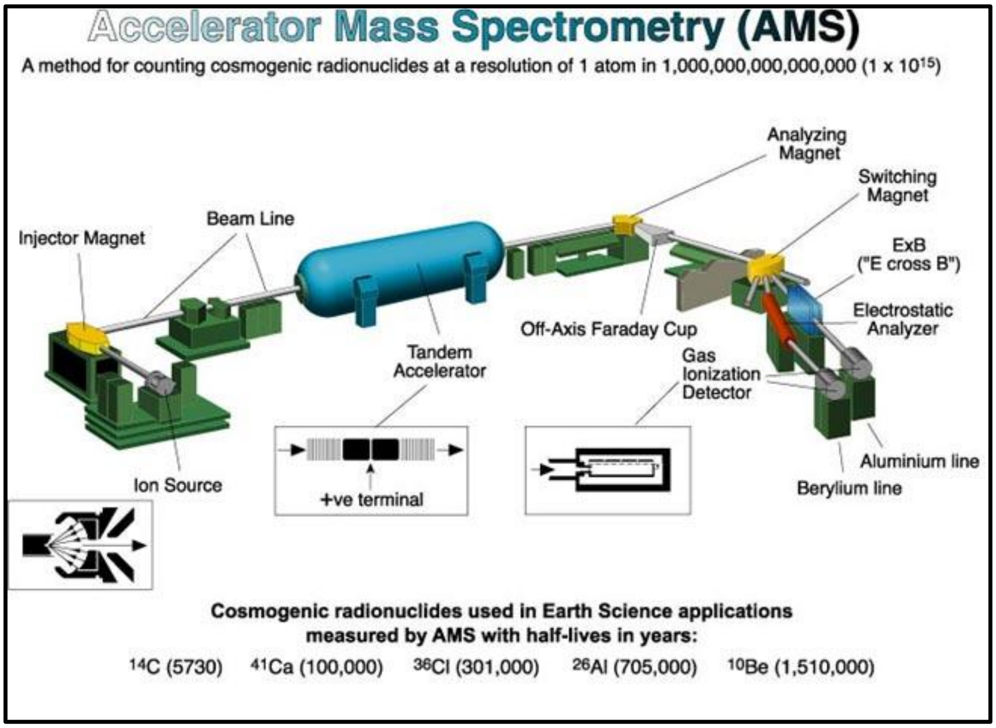

Figure 5.2: A diagram showing the ANTARES-AMS (*Australian National Tandem Accelerator for Applied Research – Accelerator Mass Spectrometer)* instrument at ANSTO (Australian Nuclear, Science and Technology Organisation (Diagram courtesy of ANSTO).

5.1.2.2. Assumptions and Corrections

SED requires several assumptions to be made in selection of samples, which include such things as the sample remaining in the same place undisturbed since placement and others discussed below and discussed in further detail by Gosse & Phillips (2001);

- Production rate – An assumption has to be made that the production rate of cosmogenic rays has been constant for the entire period the rock has been exposed. There are several components which contribute to the production rate these include;
    - Effects of geomagnetic field – due to the geomagnetic field of the earth the production rate varies with latitude. At the poles a greater quantity of cosmogenic rays are observed, than at lower latitudes.
    - Temporal variations – the variations of the production of cosmogenic rays throughout time. Due to increase cosmic activity over geological timescales the production rate may have varied. No known long term variation greater than 10 % is known, with the increase from supernovae explosions being short term events (Cerling & Craig, 1994).

- Sample coverage – the presence of any material covering the sample will shield the sample from the full exposure. Careful selection of samples free of any sediment, ice, snow is necessary.   This also includes the assumption that no previous burial has occurred by water, ice, snow, sediment etc.

- Erosion – removes material from the outer edges and surface of rocks removing some of the minerals containing the cosmogenic nuclides; this can result in a reduced exposure age. For samples in this study and area, the erosion rates are very low and are assumed to be negligible.

- Inherited concentration – when a rock has previously been exposed it will have some concentration of cosmogenic nuclides which over time will slowly decay. The sample will indicate an older than true exposure age for formation of the landform. In the field you assumed the samples have no inherited concentration, but with testing multiple sampling from a single landform and by using two different isotopes, samples with inherited concentrations can be identified.

- Lithology – some lithologies contain minerals which naturally decay to produce the same isotopes measured. This produces a high concentration of isotopes which calculates to an older than actual exposure age. For most lithologies the natural decay concentrations are so small they can be assumed to be negligible.

Corrections also have to be made to the analytical data to obtain the true exposure age for each sample. These corrections are listed below and discussed in greater detail by Gosse and Phillip, (2001);

- Atmospheric shielding – the atmospheric acts as a shield, reducing the quantity of cosmogenic rays penetrating and interacting with the rock surface. Higher altitudes receive a higher production rate due to the thinner atmosphere, lower altitudes receive lower production rate due to the thicker atmosphere. Altitude is then an important variable for the amount of shielding the atmosphere provides. Regional geological history is also a factor worth considering as if the sample has been uplifted over a period of time, it will change the quantity of cosmogenic rays interacting with the sample over the same period of time.

- Topographic shielding – topographic surrounding the sample site can change the quantity of cosmogenic rays reaching the sample. As cosmogenic rays can only penetrate through a few centimetres of rock, a nearby mountain will reduce the concentration of the rays interacting with the rocks reducing the number of spallation reactions. Topographic shielding is measured in the field from the sample site, by taking measurement every 30° for the full hemisphere.

- Sloping surfaces – sloping surfaces can shield some areas of the surface reducing the amount of cosmogenic ray interactions. The slope of the surface is recorded in the field before the sample is taken; samples taken from horizontal surfaces are preferred.

- Sample thickness – the thickness of the sample can alter the quantity of cosmogenic ray interactions. Cosmogenic ray quickly lose penetration with respect to the density of the material they are travelling through. With a typical rock density of 2.5 g/cc, at 640 mm the penetration of cosmogenic rays would be about 0.36 times the flux at the surface

(Gosse & Phillips, 2001). Therefore to obtain the best results the top few centimetres of rock is tested to provide the most accurate exposure history.

5.1.3. Applications

In-situ cosmogenic nuclide dating has become a common method of dating for geomorphic features. The applications of in-situ cosmogenic nuclide dating are sparse and can be used for more than just exposure ages. The technique has been used to assist in studies involving; erosion rates, burial histories, soil dating and also has been used in applications investigating meteoric craters, archaeology and ice cores (Cerling & Craig, 1994; Gosse & Phillips, 2001; Manz, 2002).

The use of in-situ cosmogenic nuclide dating has significantly improved dating of glacial events around the globe, some of which had been virtually impossible. Previously, glacial events were often dated by radiocarbon dating of associated carbon sources (e.g. algae, branches, charcoal). Due to the glacial processes and the climate associated with glaciers, the presence of carbon sources are restricted and often do not directly relate to the glacial event. Thus, cosmogenic dating is especially useful in the Antarctic environment where carbon material is limited to sparse marine based sources or minor deposits of lake algae.

As discussed previously, cosmogenic nuclide dating (SED) has been used in the McMurdo Dry Valley area before (Brown et al., 1991; Brook et al., 1995) with reasonable success, allowing the chronology of glaciations to be formed for the Taylor Valley and for WAIS glaciation from Ross Sea Drifts.

5.2. Aim

This study used sixteen samples, tested for both ^{10}Be and ^{26}Al. A careful selection process had to be decided upon to allow for the best results, from the limited number of samples. To provide valid data, each feature would have at least three samples tested to allow for comparisons to be made. The aim was to select features which would provide the best results for the following two aims;

1. Create an evolution history of the area. By selecting appropriate features throughout the area, age constraints can be placed on the formation of these features. Ages can be then extrapolated to the surrounding geomorphic features.

2. To determine the timing of glacial movement throughout the area and compared it to the global glacial cycle.

5.3. Method

5.3.1. Features Selected

The geomorphic features selected to be dated were chosen to incorporate the entire field area and to provide a chronology for the more recent events. By dating features within the two main valleys a generalised timing for glacial movement could be hypothesised and age interpretations could be extrapolated onto the evolution of surrounding surfaces. All field data on samples is shown in Table 5.1.

5.3.1.1. Garwood Valley

The first features selected were the terminal moraines surrounding the Garwood Glacier (Figure 4.10 & 5.4), samples collected from this feature are named GG followed by a field number. These moraines are observed as looped moraines, surrounding the terminus of the Garwood Glacier. Observed are three well defined moraines forming from lateral moraines and extending around the terminus of the present glacial limit. The most recent moraine is adjacent to the present day Garwood Glacier with ice limit currently about 20 m away. Two other moraines sit adjacent to this moraine on the western side of the glacier. The moraine are between 5 – 20 m high and are unconsolidated (loose underfoot). They are comprised of mainly granite and gneiss boulders (less than 1 m x 1 m x 1 m), with glacial till infilling the spaces between the boulders. All moraines have a dominant crest, with steeply dipping sides. All sampled were collected from the crest to reduce the likelihood of subsequent movement since deposition. The samples were dominantly granites, collected on the western side of the moraines, as the moraine crest is more pronounced.

The Joyce Glacier lineament is seen on the northern side of the Joyce Glacier (Figure 5.4), the lineament about 200 m from the present day terminus. The lineament is a small terrace/lateral moraine which runs along the northern Garwood Valley wall, with a steep edge facing towards the centre of the valley. The flat top surface only gently dips (0 - 5°) towards the east and the laterally can be traced for 500 m. The terrace is consolidate (relatively firm under foot) with large (0.5 – 2 m) boulders protruding through the surface. The samples (all samples collected from this feature are named JG followed by a field number) were collected from boulders (0.3 – 0.7 m) of granite along the edge of this lateral moraine; this should eliminate the possibility contamination from material coming off the valley walls.

Figure 5.3: Field photograph of sample JG1, the top edge of boulder was removed. Directional arrows and sample identification numbers were written directly onto the sample before removing the sample.

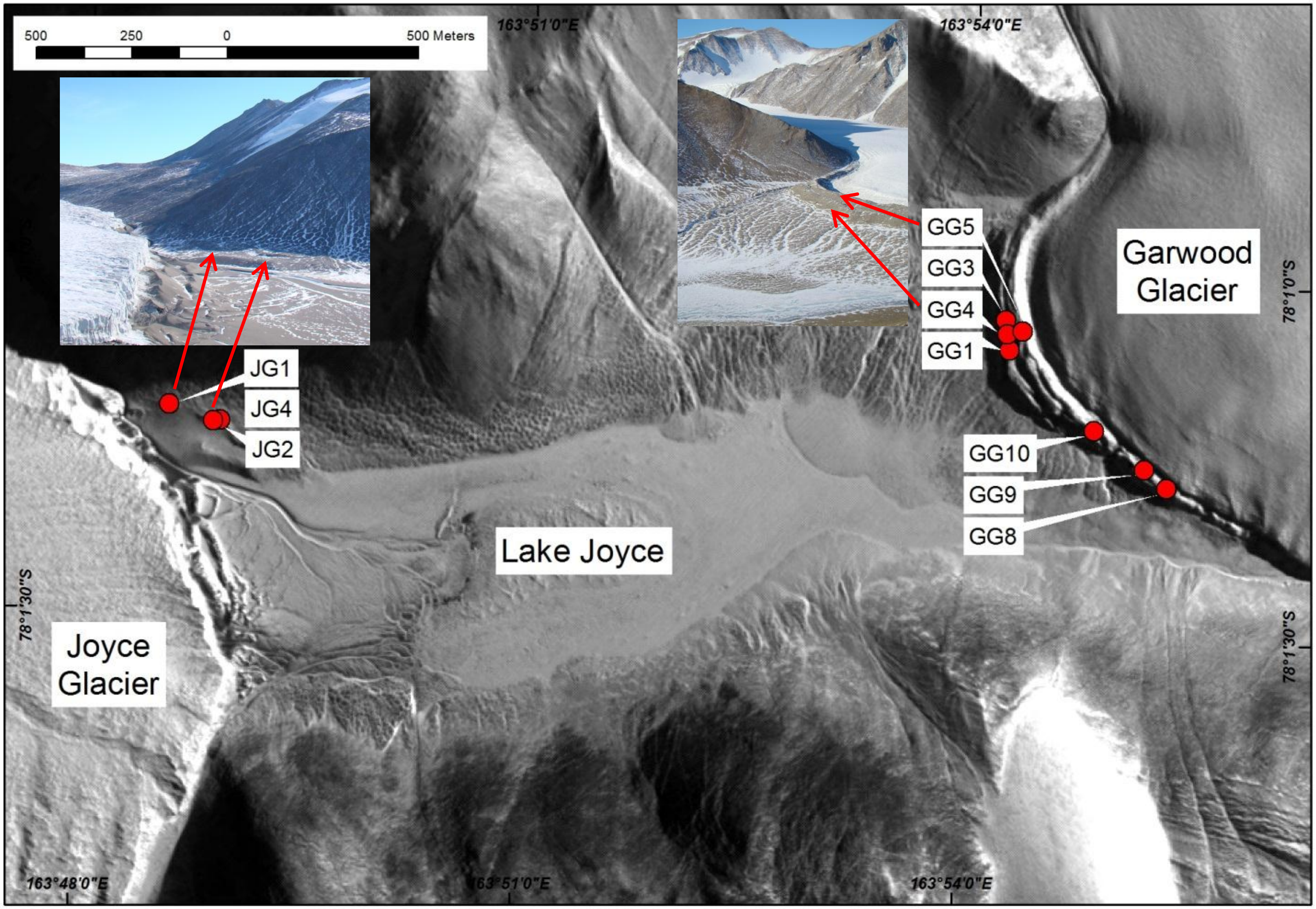

Figure 5.4: The western end of the Garwood Valley showing the locations of the SED samples, a lateral moraine near the Joyce Glacier (JG 1, JG 2 & JG 4) and from around the Garwood Glacier, an alpine glacier (GG 1, GG 3, GG 4, GG 5, GG 8, GG 9 & GG 10).

5.3.1.2. Miers Valley

Miers Glacier lateral moraine/terrace is observed wrapping around Holiday Peak from the Miers Glacier around into Adams Glacier (Figure 5.5). The samples were taken from the northern side of Holiday Peak, above the present Miers Glacier. The surface is relatively flat slightly dipping (2 - 5°) down the valley (east), conjoining with a matching surface from the Adams Glacier, on the southern side of Holiday Peak. The sediment along this feature is comprised of only granites and much smaller clast sizes (0.10 – 0.5 m) than the previous discussed features. All the samples (sample collected from this feature were named MG followed with a field number) were collected from either protruding boulders embedded into the terrace or perched cobbles upon larger boulders. The smaller clast sizes made it difficult to find samples with a good relief from the surface (0.15 – 0.45 m).

The final feature selected was a ridgeline found on the floor of the Miers Valley approximately 2 km in front of the Adams Glacier (Figure 5.5). The location and orientation of this feature was quite unusual, appearing as a moraine ridge but adjacent to Lake Miers. The feature stands about 15 m off the valley floor and is about 250 m long ridge. The ridge has been consolidated with little large material protruding from the surface. This feature like the others selected, again comprised of mostly of granitoids, however intermixed throughout are finer (pebble – coarse sand) basaltic material. The SED samples (sample were named AG followed with a field number) were collected from the most pronounced granite boulders (0.05 – 0.4 m) found along the ridgeline.

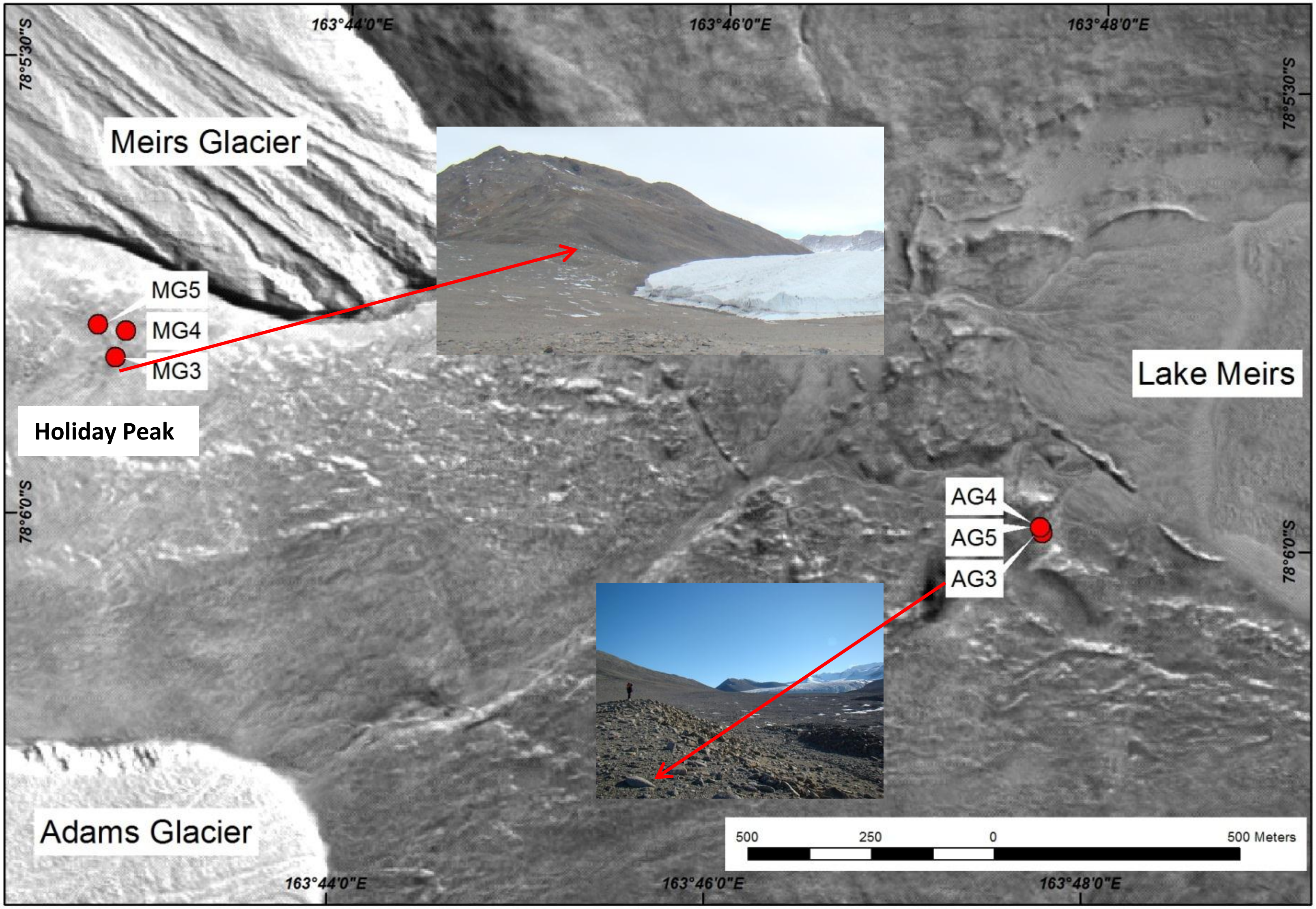

Figure 5.5: The western end of the Miers Valley showing the location of the SED samples collected from a recessional moraine around Holiday Peak (MG 3, MG 4 & MG 5) and a ridge on the valley floor near Lake Miers (AG 3, AG 4 & AG 5).

5.3.2. Collection of Samples

The sixteen samples were analysed using both ^{10}Be and ^{26}Al, which requires the refining of the target mineral (quartz). All of the landforms contained a large quantity of granitoids having been eroded from 'Granite Harbour Intrusives' within the surrounding mountains. These granites appear to have fresh surfaces and very little discolouration which suggested very little physical or chemical erosion has occurred since glacial placement.

Samples were selected by finding boulder with no evidence for movement by either being embedded into the moraine or perched upon other boulders and minimal possibility of burial by any material (e.g. sediment, ice or snow). When a sample was selected for sampling, the sample was collected from the top surface of the boulder allowing for the greatest exposure to be collected (Figure 5.6). For samples which were from smaller boulders (> 0.2 x 0.2 x 0.2 m) perched upon larger boulders the entire boulder was collected and later the top few centimetres were cut off and sampled. Larger boulder (< 0.2 x 0.2 x 0.2 m) samples were collected by using a chisel and hammer to remove the top surface. Before the sample was removed and bagged, the sample was directly labelled with the sample number and directional arrows showing which surface was at the top. The field data of the samples are summarised in Table 5.1 below.

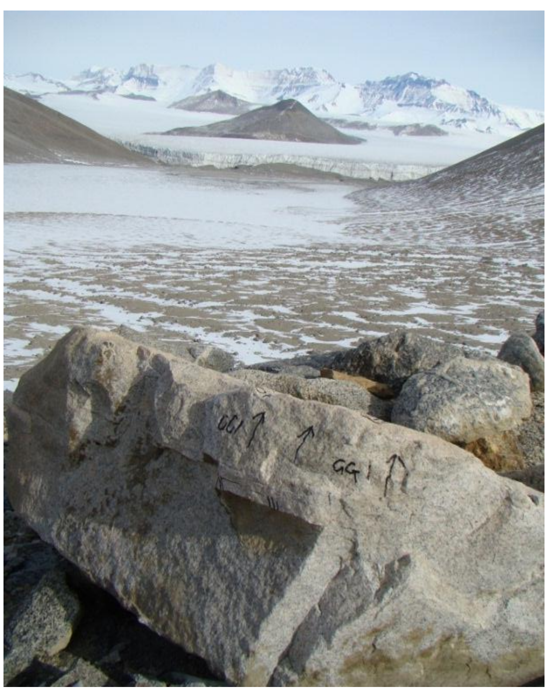

Figure 5.6: Field photograph of sample GG1, photograph shows the labelling of the samples in situ before removing section from the top of boulder.

Sample ID	Latitude	Longitude	Altitude (m)	Boulder Dimensions	Depth/ Thickness	Lithology	Topographic Shading						Comments
GG1	78.01.096S	163.54.270E	409	L - 0.90 W - 0.50 T - 0.40	50 mm	Granite	0° -	20	30° -	14	60° -	18	Sample boulder sit perched on smaller material at one end, and the other is buried about 200 mm into the moraine. The sample was taken from the angular edge along the top of the boulder, see GG1 sample photographs.
							90° -	7	120° -	4	150° -	10	
							180° -	16	210° -	11	240° -	4	
							270° -	6	300° -	14	330° -	18	
GG3	78.01.054S	163.54.242E	412	L - 0.20 W - 0.15 T - 0.20	50 mm	Pegmatite	0° -	19	30° -	12	60° -	13	The sample was collected as a whole boulder, the boulder was perched on a much larger boulder embedded into the moraine. Sample was cut from top of boulder.
							90° -	13	120° -	6	150° -	10	
							180° -	14	210° -	10	240° -	4	
							270° -	8	300° -	16	330° -	24	
GG4	78.01.074S	163.54.252E	400	L - 0.50 W - 0.45 T - 0.20	40 mm	Granite	0° -	22	30° -	14	60° -	12	Flat lying slab of granite perched on smaller material of the outer edge of the moraine. The sample was collected from one of the corners of the slabs.
							90° -	11	120° -	10	150° -	7	
							180° -	16	210° -	11	240° -	4	
							270° -	7	300° -	6	330° -	20	
GG5	78.01.069S	163.54.360E	420	L - 0.15 W - 0.15 T - 0.15	50 mm	Granite	0° -	20	30° -	14	60° -	16	The sample was collected as a whole boulder which was perched on a larger gneiss slab. The sample location is on the inner most moraine of the present glacial limit, on a well-defined crest.
							90° -	7	120° -	2	150° -	7	
							180° -	15	210° -	10	240° -	4	
							270° -	6	300° -	5	330° -	15	
GG8	78.01.284S	163.55.369E	388	L - 0.60 W - 0.40 T - 0.40	50 mm	Granite	0° -	19	30° -	10	60° -	15	Collected from the sharp crest of the inner most moraine, associated with the Garwood Glacier. The boulder is buried into the moraine with 400 mm protruding out of the surface. Sample was taken from the top surface.
							90° -	14	120° -	2	150° -	8	
							180° -	16	210° -	20	240° -	14	
							270° -	8	300° -	6	330° -	12	

GG9	78.01.259S	163.55.211E	391	L - 0.70	40 mm	Gneiss	0° -	19	30° -	12	60° -	15	This sample sit proud of the surface by 0.50 m surrounded by finer gravel-sand material. This site is towards the southwest of the moraine and the crest becomes more rounded, being wider across the top.
				W - 0.40			90° -	12	120° -	5	150° -	9	
				T - 0.50			180° -	19	210° -	16	240° -	6	
							270° -	7	300° -	6	330° -	11	
GG10	78.01 206S	163.54.864E	389	L - 0.70	50 mm	Granite	0° -	20	30° -	17	60° -	12	Flat lying granite slab perched on a mound of material. The entire thickness is exposed with finer material observed underneath it. Top surface was cut off the slab for dating.
				W - 0.30			90° -	18	120° -	8	150° -	2	
				T - 0.15			180° -	9	210° -	13	240° -	15	
							270° -	9	300° -	6	330° -	6	
JG1	78.01.208S	163.48.580E	390	L - 0.40	40 mm	Granite	0° -	15	30° -	30	60° -	27	Sample collected from and edge of the lateral moraine. The sampled boulder sits proud embedded into the moraine, with an angular edge pointing up. The sample was collected from this angular edge.
				W - 0.30			90° -	8	120° -	2	150° -	6	
				T - 0.40			180° -	11	210° -	5	240° -	8	
							270° -	9	300° -	7	330° -	6	
JG2	78.01.229S	163.48.928E	387	L - 0.30	50 mm	Granite	0° -	20	30° -	31	60° -	28	The whole boulder was collected, with the top surface subsequently removed for the analysis. The boulder was perched upon smaller pebble size material on a large granite slab. The site was along the lateral moraines surface.
				W - 0.25			90° -	16	120° -	1	150° -	12	
				T - 0.10			180° -	11	210° -	4	240° -	4	
							270° -	8	300° -	7	330° -	5	
JG4	78.01.230S	163.48.881E	390	L - 0.50	50 mm	Granite	0° -	23	30° -	30	60° -	21	Angular boulder perched on larger granite boulder buried within the edge of the lateral moraine. The top surface taken from the boulder for sampling.
				W - 0.30			90° -	2	120° -	7	150° -	11	
				T - 0.25			180° -	12	210° -	6	240° -	7	
							270° -	7	300° -	8	330° -	7	
MG3	78.05.826S	163.42.795E	317	L - 0.35	50 mm	Granite	0° -	12	30° -	7	60° -	9	Half of the original boulder was collected, later top surface cut off for analysis. Fine grained, rounded granite boulder perched on larger boulder buried
				W - 0.15			90° -	5	120° -	9	150° -	7	
				T - 0.20			180° -	2	210° -	7	240° -	10	

							270° -	9	300° -	16	330° -	14	with the moraine surface. Location above present Miers Glacier on moraine which wraps around Holiday Peak.
MG4	78.05.797S	163.42.852E	318	L - 0.35	40 mm	Granite	0° -	12	30° -	8	60° -	9	Sample collected from a boulder with the base buried in the moraine surface. The sample is from fine grained granite and with little evidence of further erosion since emplacement.
				W - 0.25			90° -	8	120° -	10	150° -	6	
				T - 0.25			180° -	2	210° -	9	240° -	5	
							270° -	14	300° -	15	330° -	9	
MG5	78.05.791S	163.42.702E	324	L - 0.50	50 mm	Granite	0° -	11	30° -	9	60° -	10	The sample was collected from the top edge of a large granite boulder buried into the moraine surface. The boulder protruded at least 300 mm from the surface, being clear from all fine material.
				W - 0.35			90° -	12	120° -	4	150° -	7	
				T - 0.30			180° -	9	210° -	7	240° -	7	
							270° -	10	300° -	14	330° -	8	
AG3	78.05.984S	163.47.727E	181	L - 0.40	50 mm	Granite	0° -	10	30° -	10	60° -	11	Sample collected from ridge adjacent to the Lake Miers margin. The sample slightly buried into the surface of the ridgeline.
				W - 0.20			90° -	8	120° -	7	150° -	1	
				T - 0.20			180° -	5	210° -	13	240° -	14	
							270° -	11	300° -	6	330° -	7	
AG4	78.05.988S	163.47.737E	184	L - 0.45	30 mm	Granite	0° -	10	30° -	11	60° -	11	Small granite slab lying on the ridgeline of a mound of a material adjacent to Lake Miers. The samples appears to be clear of fine material, and a natural fracture allow the top 30 mm to be collected for analysis.
				W - 0.25			90° -	8	120° -	7	150° -	2	
				T - 0.05			180° -	6	210° -	14	240° -	16	
							270° -	11	300° -	6	330° -	10	
AG5	78.05.982S	163.47.721	179	L - 0.15	50 mm	Granite	0° -	9	30° -	10	60° -	13	Two small boulders perched on a large granite boulder buried into the ridge adjacent to Lake Miers. Both boulders were sampled however only the granite boulder was analysed.
				W - 0.15			90° -	11	120° -	6	150° -	4	
				T - 0.10			180° -	11	210° -	15	240° -	14	
							270° -	6	300° -	11	330° -	6	

Table 5.1: Sample description and field data for the SED sample collected.

5.3.3. Preparation

On return to Christchurch, New Zealand, the samples were prepared by the Cosmogenic Preparation Laboratory at the University of Canterbury. The laboratory concentrates the quartz following processes planned and developed in collaboration with ANSTO (Australian Nuclear Science and Technology Organisation.

Initially the samples were weighed, photographed and given a laboratory number, allowing identification to field notes and photographs. The samples were crushed into fine powder using jaw crushers and pulveriser (plate grinder). The fine powder was sieved into different particle sizes, allowing the separation of the 212-500 micrometre fraction. Each sample was washed using distilled water until the liquid is clear and then dried ready for further processing.

The cleaned powder was then subjected to acid digestion to remove the remaining impurities, concentrating the quartz. The mineral lattice of quartz is very stable, which allows it to be very resistant to many acids. This property of quartz allows the sample to be washed with acid which will remove other minerals. The acid digestion process uses a series of acids; nitric acid (HNO_3), hydrochloric acid (HCl), sodium hydroxide (NaOH) and phosphoric acid (H_3PO_4) to concentrate the quartz.

After the acid digestion, a 'spike' (known concentration of the native element) was added, this means that during the analysis only the ratio of cosmogenic isotope to the known native element needed to be determined. The sample then went through an anion and cation exchange to remove the remaining unwanted ions before it was precipitated to hydroxides ($Be(OH)_3$ & $Al(OH)_3$). The precipitates were then sent for analysis by AMS.

5.3.4. Analysis

The precipitated hydroxides samples were sent to ANSTO (Australian Nuclear Science and Technology Organisation) in Sydney, Australia, for analysis. The samples were analysed by the ANTARES-AMS (*Australian National Tandem Accelerator for Applied Research – Accelerator Mass Spectrometer)* allowing the concentration of each isotope to be counted. The samples were tested for both ^{10}Be and ^{26}Al isotopes. AMS operates on the principle that charged particles that pass through a magnetic field are deflected along varied paths, due to their specific atomic mass (Olmsted & Williams, 2006).

The sample were loaded into the AMS, firstly the sample was ionised forming negatively charged particles (ions). Next ions enter an electrostatic 'tandem accelerator', which is made up of two stages and accelerates the ions. In between the two stages the ions get 'stripped' by passing through a thin latter of matter (either gas or a thin carbon foil), changing the ions from negatively charged to positively charged and any molecules break apart. These ions are accelerated and pass into the analysing magnetic field which separates the ions according to the atomic mass. The magnetic field separates the ions by deflecting the lighter ions more than heavier ions, changing the trajectories. The ions then pass into a detector, counting the abundance of each ions type.

5.4. Data

5.4.1. Data Corrections

The analytical data from the ANTARES-AMS were calibrated with the additional field information to provide a corrected total exposure concentration of the sample. The corrections were completed using both Lal (1988, 1991) and Stone (2000) corrections to provide exposure ages. This data for both the Beryllium (Be) and the Aluminium (Al) data can be seen in Table 5.2 & 5.3. Additional data use to determine scaling factors is in Appendix 3.

^{10}Be Exposure Ages (Lal/Stone Corrections)

CONSTANTS:								
^{10}Be half-life =	1.388E+06	y			Density =	2.70	g/cm^3
Lambda =	4.994E-07	y^{-1}	Err=	0	Atten length=	150.00	cm^2/g
SLHL ^{10}Be Prod Rate)=	4.600	at/g^{-s}	Err=	6.0	Atten err =	4.0	cm^2/g

Field ID	^{10}Be/9Be ASB Ratio (1E-15)	Absolute Error	Percentage Error	Sample Mass (g)	Carrier Mass (mg)	^{10}Be Conc (atoms/g-Q) (1E6)	^{10}Be Conc Err (atoms/g-Q) (1E6)	Site Specific Prod. Rate (at/g/yr)	Min ^{10}Be Age (ka)	QuaD Error (ka)	Max Erosion rate (mm/ka)	Erosion rate error (mm/ka)	^{10}Be conc SLHL (1E6)	^{10}Be Conc Error SLHL (1E6)	^{10}Be fraction of saturation SLHL
GG1	1133.70	11.40	1.01	101.49	0.393	0.2934	0.0072	**8.428**	**35.12**	**2**	15.68	1.01	0.155	0.004	0.017
GG3	1302.60	12.00	0.92	98.23	0.412	0.3651	0.0088	**8.269**	**44.65**	**3**	12.30	0.81	0.194	0.005	0.021
GG4	292.70	10.40	3.55	47.46	0.489095	0.2016	0.0085	**8.184**	**24.79**	**2**	22.28	1.66	0.108	0.005	0.012
GG5	783.30	17.20	2.20	80.58	0.452525	0.2940	0.0092	**8.302**	**35.73**	**2**	15.41	1.07	0.154	0.005	0.017
GG8	1768.40	20.20	1.14	102.05	0.417565	0.4836	0.0121	**8.038**	**61.09**	**4**	8.96	0.60	0.262	0.007	0.028
GG9	483.80	15.10	3.12	79.4	0.474375	0.1932	0.0074	**8.145**	**23.86**	**2**	23.15	1.67	0.104	0.004	0.011
GG10	505.60	18.00	3.56	80.19	0.424	0.1785	0.0075	**8.077**	**22.23**	**2**	24.86	1.86	0.097	0.004	0.010
JG1	371.70	10.50	2.82	46.5	0.47357	0.2530	0.0091	**8.067**	**31.61**	**2**	17.44	1.25	0.137	0.005	0.015
JG2	735.40	12.90	1.75	80.42	0.384	0.2349	0.0067	**8.125**	**29.12**	**2**	18.94	1.26	0.127	0.004	0.014
JG4	430.90	16.10	3.74	54.66	0.47495	0.2502	0.0109	**8.075**	**31.23**	**2**	17.65	1.34	0.135	0.006	0.015
MG3	669.00	18.30	2.74	80.1	0.497375	0.2776	0.0098	**7.549**	**37.12**	**3**	14.83	1.05	0.161	0.006	0.018
MG4	615.20	31.20	5.07	80.46	0.403765	0.2063	0.0114	**7.625**	**27.24**	**2**	20.25	1.69	0.120	0.007	0.013
MG5	509.10	16.80	3.30	79.08	0.47771	0.2055	0.0082	**7.606**	**27.21**	**2**	20.28	1.49	0.119	0.005	0.013
AG3	472.80	10.30	2.18	100.62	0.40802	0.1281	0.0040	**6.567**	**19.61**	**1**	28.19	1.93	0.085	0.003	0.009
AG4	237.20	8.60	3.63	80.26	0.47265	0.0934	0.0040	**6.499**	**14.42**	**1**	38.39	2.91	0.062	0.003	0.007
AG5	537.00	10.30	1.92	99.99	0.418485	0.1502	0.0044	**6.495**	**23.26**	**1**	23.74	1.62	0.100	0.003	0.011

Table 5.2 – ^{10}Be data with Lal (1988, 1991) and Stone (2000) corrections

^{26}Al Exposure Ages (Lal/Stone Corrections)

CONSTANTS:	**^{26}Al half-life =**	7.010E+05	**y**			**Density =**	3	**g/cm^3**
	Lambda =	9.888E-07	**y^{-1}**	**err=**	0.0	**Atten length=**	150	**cm^2/g**
SLHL ^{26}Al Prod Rate)=		31.100	**at/g^{-s}**	**err=**	6.0	**Atten err =**	4	**cm^2/g**

Field ID	^{26}Al/^{27}Al ABS Ratio (1E-15)	Absolute Error	Percentage Error %	Sample Mass (g)	Carrier Madd (mg)	Al ppm	Al Error ppm	Native Al (mg)	Total Al (mg)	^{26}Al Conc (atoms/g-Q) (1E6)	^{26}Al Conc Err (atoms/g-Q) (1E6)	Site Prod Rate (at/g/yr)	Min ^{26}Al Age (ka)	QuaD Error (ka)	Max erosion rate (mm/ka)	Erosion rate error (mm/ka)	^{26}Al conc SLHL (1E6)	^{26}Al Conc Error SLHL (1E6)	^{26}Al fraction of saturation SLHL
GG1	1215.55	135.2	11.13	101.49	0	89.653	2.0	9.10	9.10	2.43	0.29	56.98	43.62	5.82	12.46	1.70	1.29	0.16	0.04
GG3	600.94	127.7	21.25	98.23	0	243.904	2.0	23.96	23.96	3.27	0.71	55.91	60.28	13.33	8.94	2.20	1.74	0.38	0.06
GG4	132.51	48.2	36.38	47.46	0	334.545	2.0	15.88	15.88	0.99	0.36	55.33	18.05	6.44	30.51	12.02	0.53	0.19	0.02
GG5	833.97	227.1	27.23	80.58	0	72.001	2.0	5.80	5.80	1.34	0.37	56.13	24.17	6.55	22.71	6.84	0.70	0.19	0.02
GG8	2033.05	225.2	11.08	102.05	0	74.704	2.0	7.62	7.62	3.39	0.41	54.34	64.39	8.43	8.36	1.18	1.84	0.22	0.06
GG9	990.59	121.7	12.29	79.4	0	69.423	2.0	5.51	5.51	1.54	0.20	55.07	28.27	3.96	19.38	2.85	0.83	0.11	0.03
GG10	269.31	60.3	22.41	80.19	0	205.060	2.0	16.44	16.44	1.23	0.28	54.61	22.83	5.19	24.06	6.01	0.67	0.15	0.02
JG1	531.21	105.9	19.93	46.5	0	147.753	2.0	6.87	6.87	1.75	0.36	54.54	32.64	6.70	16.75	3.80	0.95	0.19	0.03
JG2	726.83	108.4	14.91	80.42	0	98.278	2.0	7.90	7.90	1.59	0.25	54.93	29.45	4.79	18.59	3.21	0.86	0.14	0.03
JG4	423.33	141.2	33.36	54.66	0	220.733	2.0	12.07	12.07	2.09	0.70	54.60	38.95	12.87	13.99	5.19	1.13	0.38	0.04
MG3	1210.57	217.5	17.97	80.1	0	76.747	2.0	6.15	6.15	2.07	0.38	51.04	41.48	7.86	13.12	2.72	1.20	0.22	0.04
MG4	720.22	136.0	18.89	80.46	0	95.468	2.0	7.68	7.68	1.53	0.30	51.55	30.22	6.00	18.11	3.84	0.89	0.17	0.03
MG5	712.94	155.0	21.74	79.08	0	85.282	2.0	6.74	6.74	1.36	0.30	51.42	26.74	5.95	20.50	4.98	0.78	0.17	0.02
AG3	264.24	85.3	32.27	100.62	0	84.153	2.0	8.47	8.47	0.50	0.16	44.40	11.24	3.56	49.14	17.15	0.33	0.11	0.01
AG4	412.68	95.9	23.24	80.26	0	69.328	2.0	5.56	5.56	0.64	0.15	43.94	14.64	3.38	37.67	9.78	0.42	0.10	0.01
AG5	471.51	86.7	18.39	99.99	0	131.607	2.0	13.16	13.16	1.39	0.26	43.91	32.05	6.10	17.06	3.62	0.92	0.18	0.03

Table 5.3 – ^{26}Al data with Lal (1988, 1991) and Stone (2000) corrections

5.4.2. Data Interpretation

To provide valid data two isotopes (^{10}Be & ^{26}Al) have been used to test each sample, this allows a comparison of the estimated ages to be made, which help define the exposure history of the sample (A table of ^{10}Be and ^{26}Al ages can be found below in Table 5.4). When comparing the estimated ages from the isotope data several outcomes can be identified, these are;

- If the two ages are within error of each other, it suggests that the age is a true exposure age. This true exposure age means the sample has either been exposed to a single exposure event or the samples has been buried for a significant period to erase any pre-existing cosmogenic isotopes.

- If the ^{26}Al age < ^{10}Be age, it suggests the sample has been subjected to a complex exposure history. The half-life of the two isotopes are different with ^{26}Al decaying faster than ^{10}Be. If the sample had been previously exposed and then buried the isotopes would start to decay. The ^{26}Al isotope would decay quicker reducing the concentration, therefore the reducing the measured ^{26}Al concentration. The sample could then be reworked and deposit again on the surface, having an inherited unbalanced concentration of the two isotopes.

- If the ^{26}Al age > ^{10}Be age, this suggests there could be a problem with the technique or inherited isotope. The half-life of ^{26}Al is short than the ^{10}Be half-life therefore it should be impossible for ^{26}Al so show an older age, however if rock contains a proportion of naturally occurring ^{26}Al then this could complicate the analysis. ^{26}Al is difficult to refine in the laboratory, therefore we could assume that the ^{10}Be is correct/true exposure age.

Sample	^{10}Be Age (ka)	^{10}Be Age Error (ka)	^{26}Al Age (ka)	^{26}Al Age Error (ka)
GG1	35.12	2.25	43.62	5.82
GG3	44.65	2.82	60.28	13.33
*GG4	24.79	1.76	18.05	6.44
*GG5	35.73	2.34	24.17	6.55
GG8	61.09	3.86	64.39	8.43
GG9	23.86	1.65	28.27	3.96
GG10	22.23	1.57	22.83	5.19
JG1	31.61	2.13	32.64	6.70
JG2	29.12	1.89	29.45	4.79
JG4	31.23	2.23	38.95	12.87
MG3	37.12	2.51	41.48	7.86
MG4	27.24	2.17	30.22	6.00
MG5	27.21	1.90	26.74	5.95
*AG3	19.61	1.28	11.24	3.56
AG4	14.42	1.01	14.64	3.38
^AG5	23.26	1.49	32.05	6.10

Table 5.4: ^{10}Be and ^{26}Al ages of all samples. *sample with ^{26}Al age < ^{10}Be suggesting a complex exposure history. ^ Sample with ^{26}Al age > ^{10}Be, an unusual result.

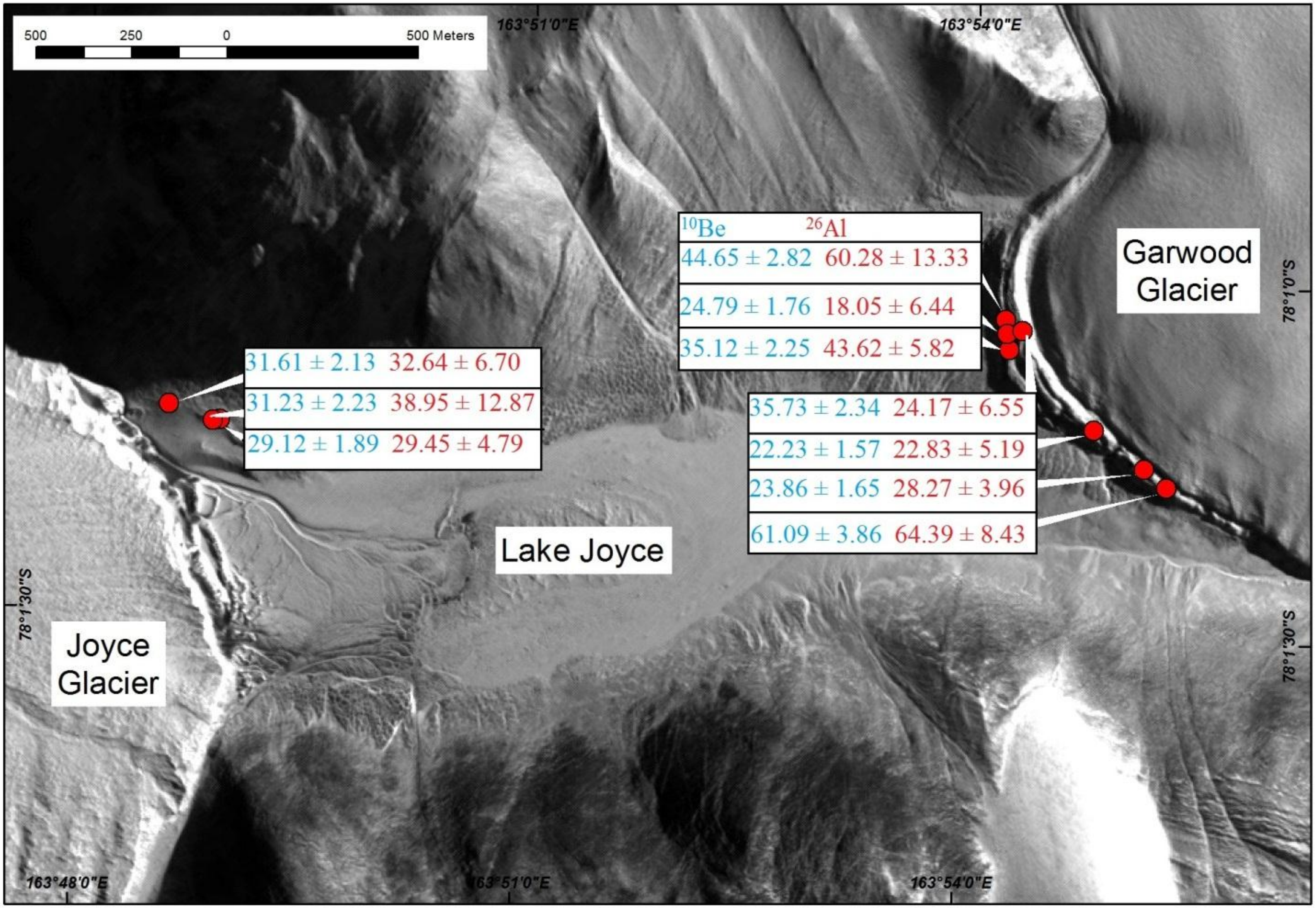

Figure 5.7: The western end on the Garwood Valley with the SED ages. The ages in red (left) are ^{10}Be ages and the blue (right) are ^{26}Al ages.

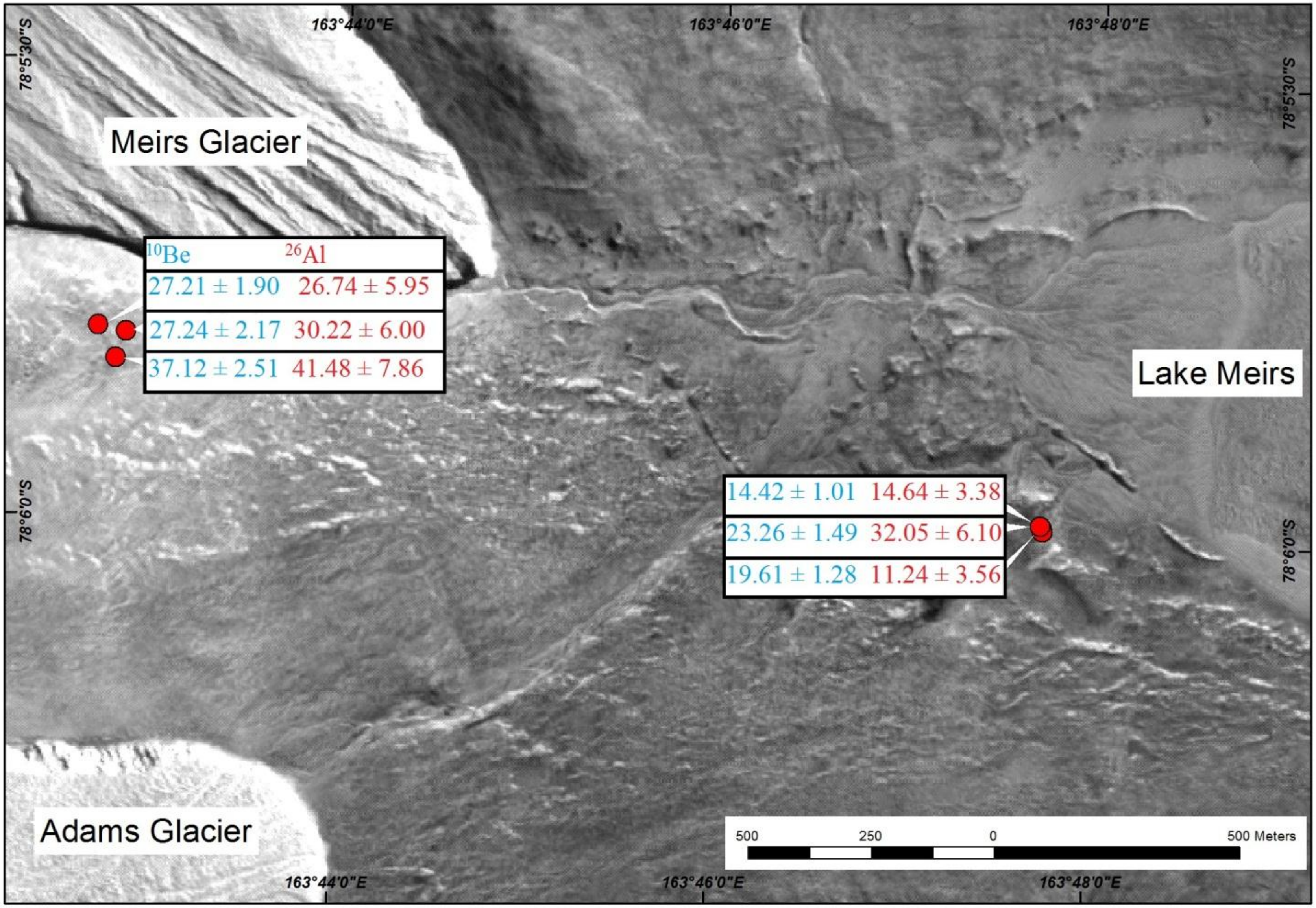

Figure 5.8: The western end of the Miers Valley with the SED ages. The ages in red (left) are the ^{10}Be ages and the blue (right) are ^{26}Al ages.

5.4.2.1. Data Analysis

To analysis the data four graphs have been plotted to determine any common relationships; banana plot (burial history), age verses altitude, $^{26}Al/^{10}Be$ ratio verses height above surface and $^{26}Al/^{10}Be$ ratio verses clast size of sample (Figure 5.9, 5.10, 5.11 & 5.12).

The banana plot (Figure 5.9) suggests that at least two of the samples (GG5 and AG3) have been buried for a considerable amount of time. All the other samples are above the exposure/erosion curves which should not occur, because of the young age given by the samples the banana curve is very narrow at this end, and most of the sample cross into the curve within the error of margin.

To determine if a relationship exists between the exposure age and the altitude the sample were collected from against the exposure ages, Figure 5.10 was plotted. It has to be noted that the two valleys have a 250 m difference in altitude of the valley floor. The Garwood Valley has a steeper gradient from McMurdo Sound compared to the shallow incline of the Miers Valley. The graph indicates Adam Glacier (AG) samples display the youngest ages (11 – 32 ka) and are at the lowest altitude, which displays the normal expected behaviour. Moving further up the valley walls the Miers Glacier (MG) samples show an older age of 27 – 41 ka. From the Garwood Valley, GG & JG samples were all from the same altitude (387 – 412 m), yet display a range of ages from 22 ka up to 64 ka.

The $^{26}Al/^{10}Be$ ratio is compared to the height of the surface the sample was collected (Figure 5.11). This was to determine if the samples where the ^{26}Al and ^{10}Be ages do not correlate, were possibly due to the samples being not very pronounced and therefore having the possibility of being covered by other material. If any material covers the samples it prevents cosmogenic rays from reaching the sample, also the cosmogenic isotopes decay therefore reducing the concentration. The graph suggests there is no correlation between the height in which the sample is protruding out of the surface and the potentially buried samples.

The final plot (Figure 5.12) uses $^{26}Al/^{10}Be$ ratio compared to the size of the boulder the sample was taken from. The size has been justified by the largest dimension of the boulder, to determine whether the size of the boulder is a factor to the likelihood of it remaining stable in place since emplacement. The data is inconclusive whether a relationship could be proven into the size of the boulders and the stability, the samples with non-matching age estimates are of varying sizes. Some samples (MG5, JG1 & JG2) with ^{10}Be and ^{26}Al ages correlating very well, are less than 400 mm in maximum length.

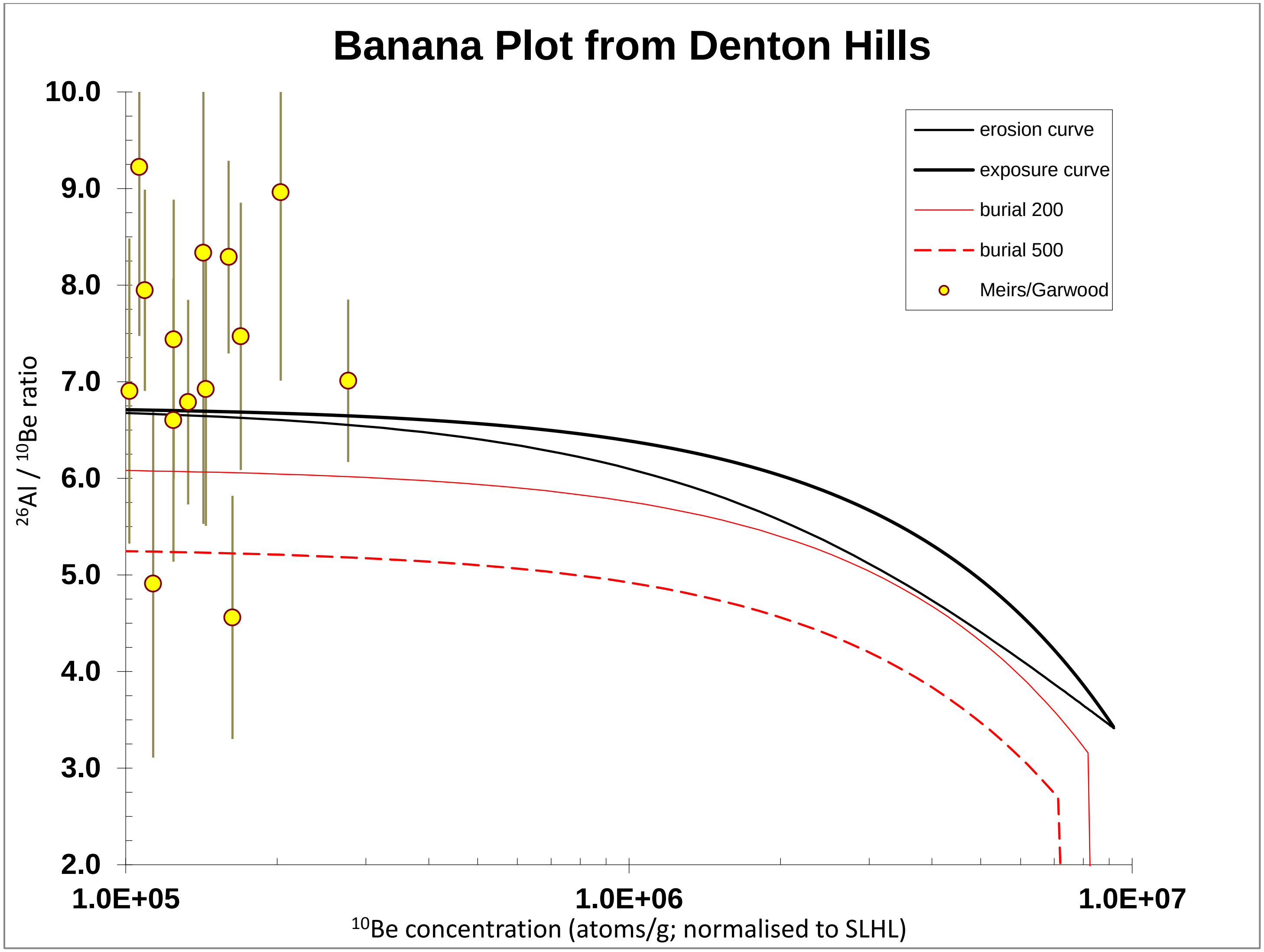

Figure 5.9: Banana Plot, comparing the ^{10}Be concentration and the ^{26}Al/^{10}Be ratio. The exposure curve and burial curves are shown to determine the history of the samples.

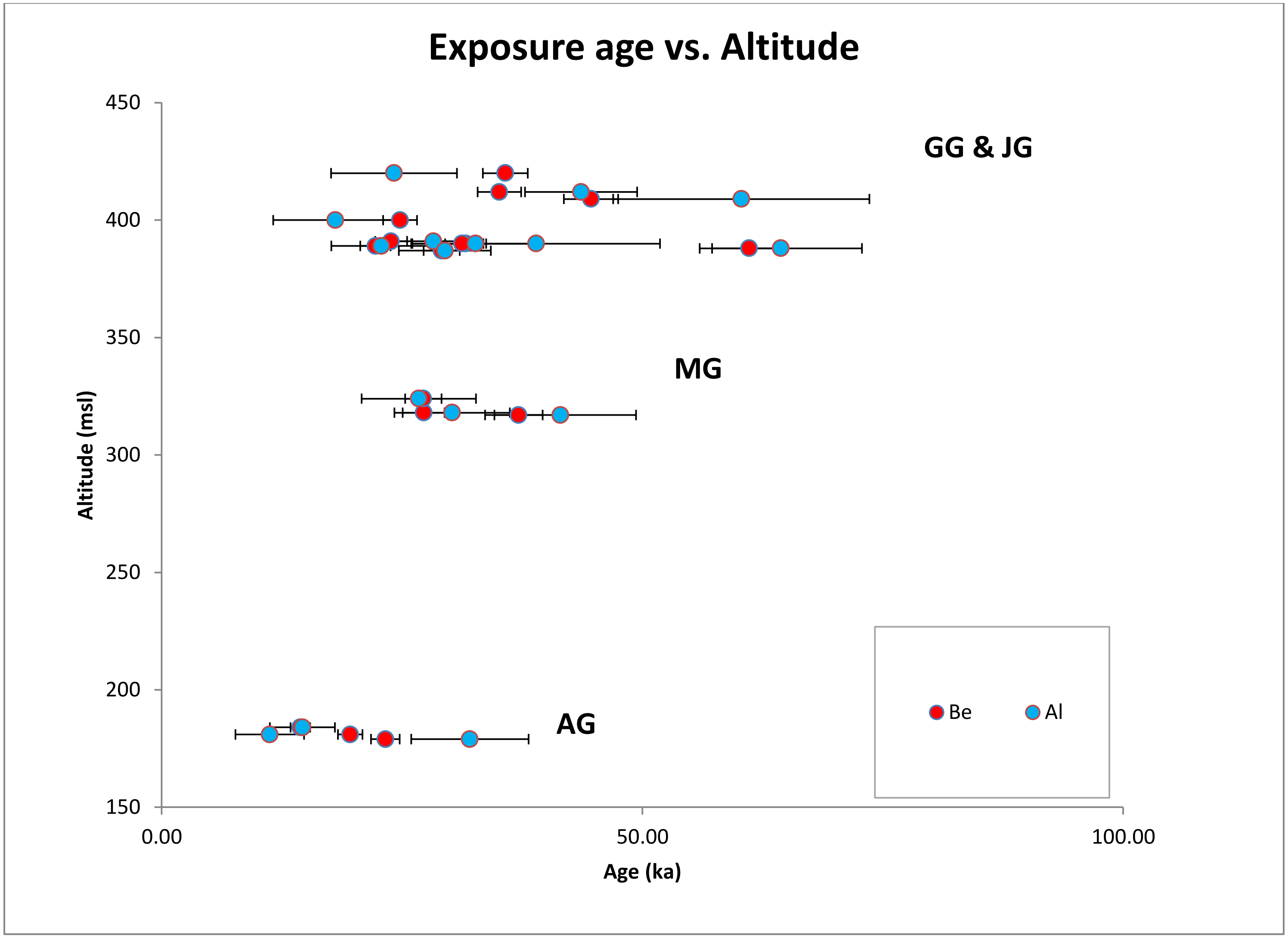

Figure 5.10: A graph comparing the ^{10}Be and ^{26}Al ages against the Altitude in which the sample was collected.

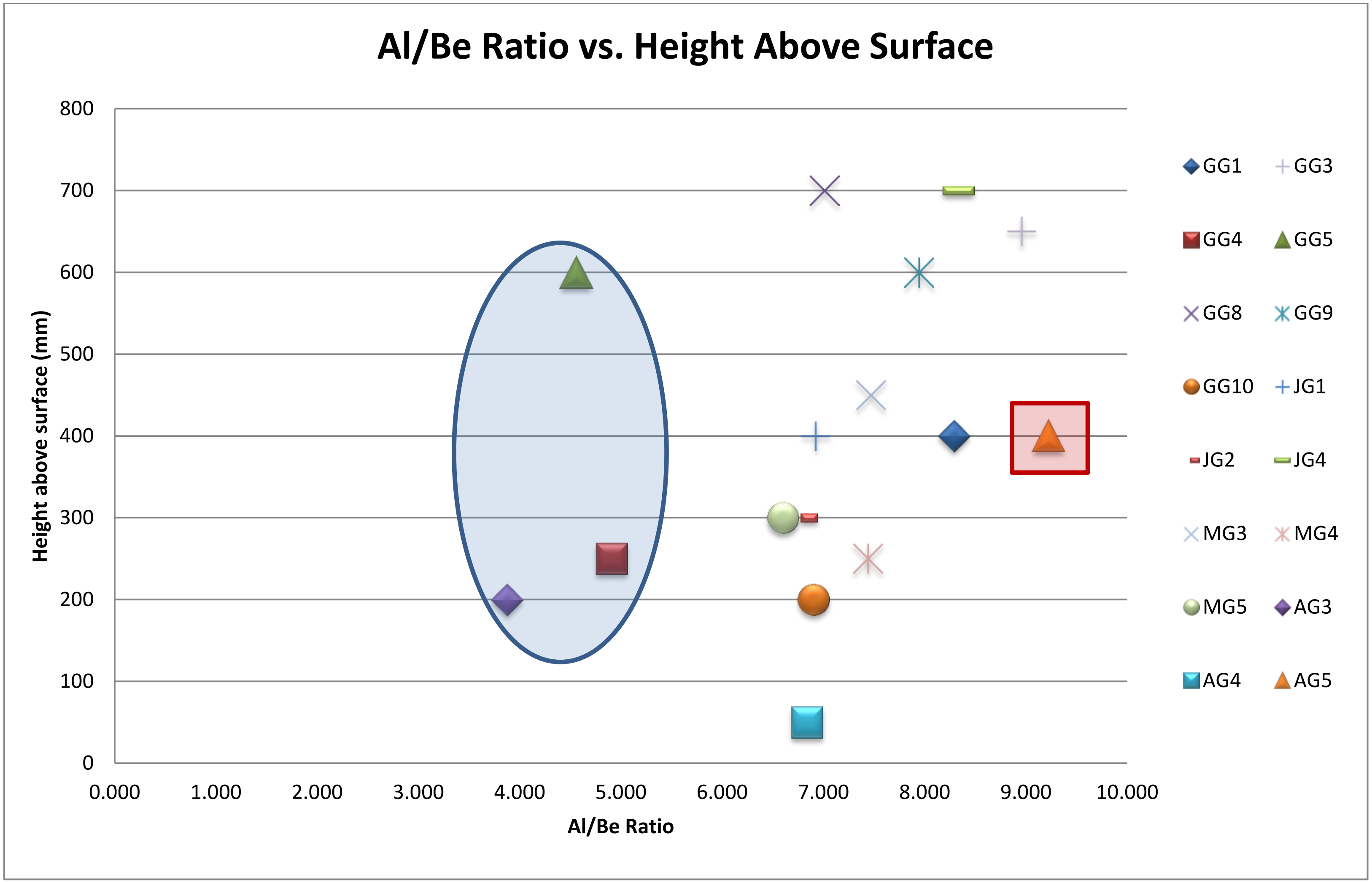

Figure 5.11: A graph expressing the $^{26}Al/^{10}Be$ ratio against the height above the surface the sample was taken. In the blue circle are the samples showing a ^{26}Al age < ^{10}Be age. The red square shows the sample with ^{26}Al age > ^{10}Be age.

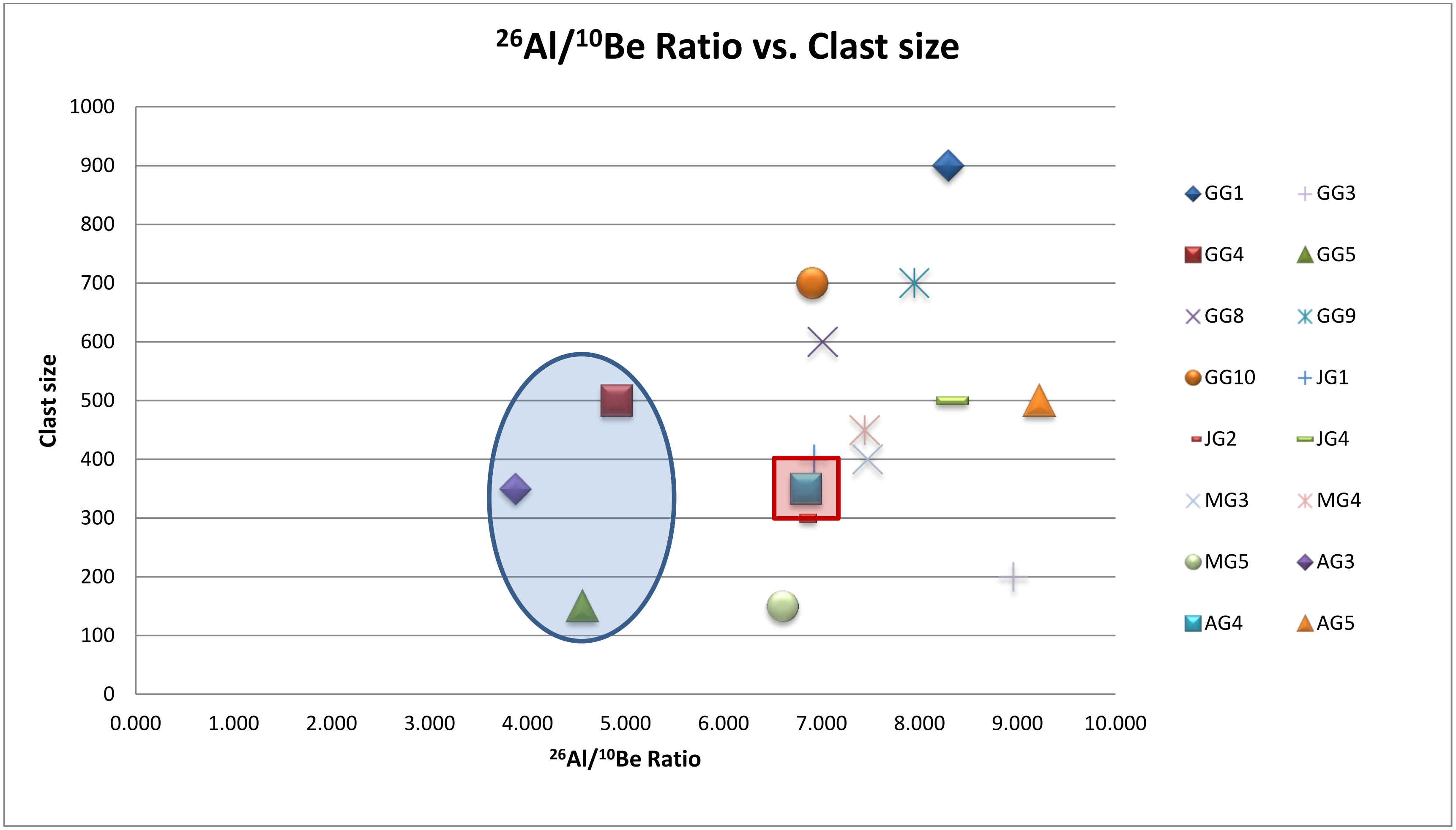

Figure 5.12: ^{26}Al/^{10}Be Ratio verses the clast size of the sample. The clast size has been determined by being the maximum length of the clast. Samples within the blue circle have a ^{26}Al age less than ^{10}Be, and the sample within the red square have a ^{26}Al age greater than ^{10}Be age.

5.4.2.2. Age interpretation

Garwood Glacier Moraine Ridges

The Garwood moraines as discussed above are relatively young looking moraines with sharp crests and unconsolidated (loose under foot). The data has provided a range of results discussed below;

The outer moraine (3)

Sample No.	^{10}Be age (ka)	^{26}Al age (ka)	Interpretation
GG1	35.12 ± 2.25	43.62 ± 5.82	The ages are close to overlapping within the error of margin. This would suggest an age of around 38 ka. The ^{26}Al age being greater than the ^{10}Be age could suggest there is a problem with the ^{26}Al dating.
GG3	44.65 ± 2.82	60.28 ± 13.33	The ages have a large discrepancy between the ages however with the ^{26}Al age having such a large margin of error associated with it the ages can overlap. This would suggest an age of about 47 ka.
GG4	24.76 ± 1.76	18.05 ± 6.44	These ages show a correlation between the margins of error however because the ^{26}Al age < ^{10}Be age, it would suggest that the sample has undergone a complex exposure history. It shows evidence of being previously exposed before getting buried and reworked.

Table 5.5: Interpretation of ages for Moraine 3 Garwood Glacier

Moraine 1 – Inner most moraine

Sample No.	^{10}Be age (ka)	^{26}Al age (ka)	Interpretation
GG5	35.73 ± 2.34	24.17 ± 6.55	These ages suggest the sample has undergone a complex exposure history being buried and re worked. The sample is from a small rounded boulder perched upon a larger granite slab embedded in the moraine crest. The physical characteristics of the sample could confirm the reworked suggestion.
GG8	61.09 ± 3.86	64.39 ± 8.43	This sample shows a good correlation between the two isotopes, suggesting the age is the true exposure age. The location of this sample would expect a younger age than the dating has provided.
GG9	23.86 ± 1.65	28.27 ± 3.96	These ages have a good correlation between the two isotopes, allowing the assumption that the ages are true exposure ages. ^{26}Al age > ^{10}Be suggesting there could be a slight discrepancies with the ^{26}Al age.
GG10	22.23 ± 1.57	22.83 ± 5.19	These ages are correlated, both agreeing with an age of 22 ka, this therefore is the true exposure age.

Table 5.6: Interpretation of ages for Moraine 1 Garwood Glacier

The exposure ages discussed above show the inner most moraine (Moraine 1) has a good correlation between two samples suggesting a formation age of 22 ka, this correlates with the LGM (Table 5.6). Yet the outer moraine's (Moraine 3) samples have provided varied results with one sample having to be disregarded due to large discrepancies between the two isotopes, suggesting a complex history of burial and reworking (Table 5.5). The remaining samples has provided an age range of between 38 - 47 ka and is only suggestive as the error of margins associated with these sample are considerably large.

Joyce Glacier Lineament

Sample No.	^{10}Be age (ka)	^{26}Al age (ka)	Interpretation
JG1	31.61 ± 2.13	32.62 ± 6.70	Both isotopes correlate in the exposure ages, suggesting a true exposure age of 32 ka.
JG2	29.12 ± 1.89	29.45 ± 4.79	Both isotopes agree with an exposure age of 29 ka, suggesting that this is the true exposure age of the sample.
JG3	31.23 ± 2.23	38.95 ± 12.87	These ages agree within the margin of error suggesting the expansion of the Joyce Glacier about 31 ka, an true exposure age.

Table 5.7: Interpretation of SED ages for the Joyce Glacier lineament.

The samples collected from around the Joyce Glacier have provided excellent results with both isotopes agreeing within the sample and all the samples suggesting the Joyce Glacier was expanded around 30 ka (Table 5.7). This expansion is pre LGM (Last Glacial Maximum) suggesting that during the LGM the Joyce Glacier did not considerably expand.

Miers Glacier Lineament

Sample No.	^{10}Be age (ka)	^{26}Al age (ka)	Interpretation
MG3	37.12 ± 2.51	41.48 ± 7.86	A good correlation between the two isotopes provides a true exposure age of 37-41 ka.
MG4	27.24 ± 2.17	30.22 ± 6.00	A good correlation between the two isotopes provides a true exposure age of 27-30 ka
MG5	27.21 ± 1.90	26.74 ± 5.95	A very good correlation between the isotopes provides a true exposure age of about 27 ka.

Table 5.8: Interpretation of SED ages for Miers Glacier lineament.

The samples collected from the terrace above the Miers Glacier have provided good correlation and a suggested age of formation about 27 ka (Table 5.8). This age is at the start of the LGM, suggesting the glacier had expanded before the LGM and has been in retreat since the start of the LGM.

Adams Glacier Ridge

Sample No.	^{10}Be age (ka)	^{26}Al age (ka)	Interpretation
AG3	19.61 ± 1.28	11.24 ± 3.56	The ^{26}Al age < ^{10}Be age suggesting the sample has been exposed to a complex exposure history. The sample has most likely have been previously exposed, buried and reworked. This would leave an inherited concentration of cosmogenic isotopes (more ^{10}Be as the half-life is smaller)
AG4	14.42 ± 1.01	14.64 ± 3.38	This sample the two isotopes agree of an exposure age of 14.5 ka, meaning this is a true exposure age.
AG5	23.26 ± 1.49	32.05 ± 6.10	No correlation between the two isotopes is seen in this sample and the ^{26}Al age > ^{10}Be age which suggests there could be a problem with the ^{26}Al dating.

Table 5.9: Interpretation of SED ages for Adams Glacier ridge.

The Adams Glacier feature is located on the floor of the Miers Valley, near the present day lake (Refer to Figure 5.8). The feature is associated with the 'lake-ice conveyor' (Figure 4.15) deposits that are formed when the valley is occupied by a proglacial lake (Clayton-Greene, 1986). The proglacial lake formed when the mouths of the valleys were dammed by ice in McMurdo Sound, and the lake rises. The deposit gets transported along the ice surface of the lake into the valley further than the glacier reached (See Chapter 4). These dates provided by the SED of this feature suggest the ridgeline was formed during the LGM, suggesting a proglacial lake occupied the valley during this period (Table 5.9).

5.4.2.3. Geomorphic Interpretations

The SED data have provided age constrains to the formation and evolution of the landforms within the Denton Hills area. The dates suggest that the larger EAIS fed glaciers did not expand significantly during or since LGM.

The Garwood Glacier moraines have provided good information about the age of the surrounding surfaces and the expansion of alpine style glaciers. The data provides an estimated formation of the inner (closest to present glacial limit) moraine of being approximately 23 ka. This age would be appropriate for a LGM expansion. The outer moraine (3) has provided a more complex set of dates with samples showing both reworking and true exposure histories. The sample which could be considered to have a true exposure age would be GG3, with an age of 47 ka. This age means during the LGM the Garwood Glacier did not expand greater than this margin.

The Joyce Glacier lineament has provided the best correlation between samples, with all three samples agreeing with an estimated age of about 30 ka. The formation of this terrace/lateral moraine is estimate to have formed just before the start of the LGM. This could be an important discovery, as two theories can be determined from it; ***a.*** that since 30 ka most of the western end of Garwood Valley has been exposed. There is evidence in the eastern end of the valleys for occupation of ice from McMurdo Sound during the LGM. And ***b.*** the Joyce Glacier did not expand considerably during the LGM. The close proximity of the terrace to the present day glacial limit suggests there has been little change in the ice volume since this period.

Samples for the Miers Glaciers lineament around Holiday Peak have provided a good correlation of two samples for an age about 27 ka. This age corresponds directly with the onset of the LGM. This suggests the Miers Glacier was at a greater volume prior to the LGM and has not been larger since the LGM. This behaviour could potentially show a retreat before the LGM and with further dating be evidence to support the 'out-of-phase' hypothesis from Marchant et al., 1994 formed from the behaviour from the Taylor Glacier in the Arena Valley.

The Adams Glacier feature along the floor of the Miers Valley has provided the most interesting SED data to interoperate. The data has shown sample with complex exposure histories and sample with simple singular exposure history. The mix of data suggests the deposit has been created from reworked material, probably having being deposited from several different sources before being finally deposited in the ridge observed today. The geomorphic interpretation from the field mapping, suggests the deposition of the feature was associated with a large pro-glacial lake which would have occupied the valley during and for some time after the LGM. The true exposure age of AG2 has provided an age of about 14.5 ka, which would closely associate with the end of the LGM and also with radiocarbon dates from Denton& Hughes (1981), Hendy (2000a,b) etc. This date is likely to be the last event to have significantly changed the appearance of the Miers Valley.

6. Discussion

6.1. Introduction

The focus of this research was to study and determine the evolutionary history of the Denton Hills area from bedrock geology to the landscape. To determine the evolution history, geological and geomorphological data were investigated by creating new maps showing the spatial distribution of geological units and geomorphic features. Surface exposure dating further extended the research providing ages for the timing of specific glacial events. The timing of glacial events allows a chronology to be assigned to the evolution of the landscape and compare them to the global glacial cycle. Below I describe the original research contributions resulting from my thesis and how they improved upon previous knowledge of the area.

6.2. Geological Evolution

6.2.1. Introduction

Although the geological units throughout the McMurdo Dry Valleys have been extensively studied, the spatial distribution was relatively unidentified. This research further defined the spatial distribution from the original geological map produce by Blank et al., 1963.

6.1.2. Spatial Distribution of Geological Units

6.1.2.1. Koettlitz Group Metasediments

The distribution of the Koettlitz Group Metasediments has been further defined by this research from the original distribution of Blank et al., (1963). The northern wall of the Garwood Valley is dominated by the Salmon Marble, bordered by gneiss on the eastern and western sides.

6.1.2.2. Granitoids

The dominant lithology throughout the study area is granitoids associated with the "Granite Harbour Intrusive Complex". The granitoids have been mapped as an individual unit for simplicity, however detail geochemical studies (Smillie, 1989; Worley, 1992) have shown that there have been multiple phases of emplacement. The granitoids dominate the centre of the study area, incorporating the occasional Koettlitz metasediment. My geological map indicates that the granitoids become more dominant to the west completely covering Holiday Peak.

6.1.2.3. Basaltic Dykes

Basaltic dykes scatter throughout the field area, being more abundant on the southern wall of the Miers Valley and on the north-eastern wall of the Garwood Glacier. Many of the dykes outcrop as only 0.1 – 0.5 m thick sub vertical traces, but larger dykes 1 – 10 m have been mapped. The mapped dykes in the southern wall of the Miers Valley trend 035 – 060° (NE – SW). This pattern has been observed by Jones (1996) in Hidden Valley (next valley to the south) but was not previously identified here prior to this study. The dykes have been suggested to have had intruded during a period of extension associated with the rifting of the Ross Sea.

6.1.3. Geological Interpretation

The bedrock geological evolution (for map see Appendix Sheet 1) can be separated into three distinct events;

1. *Formation of the Koettlitz Metasediments*. This would have occurred at some depth metamorphosing the original marine sediments. Other research has shown that metasediments have undergone several phases of deformation. The data for the

geological map indicated lineations within the metasediments are folded, indicating further deformation of the unit during ductile formation.

2. *Emplacement of the 'Granite Harbour Intrusive Complex'*. The granitoids from this complex are seen dominating the study area, intruding through the metasediments. Contacts between the two units show little sign of alteration, from either thermal or chemical, suggesting that the metasediments were still at a considerable depth and temperature when emplacement occurred. Within some of the granitoids a lineation can be observed, often matching the lineation's of the metasediments, suggesting deformation occurred after the emplacement of these granitoids.

3. *Emplacement of mafic intrusions*. Throughout the area mafic intrusions are seen either as basaltic dykes or a larger mafic intrusion near Penance Pass. These intrusions are associated with the Ross Sea rift and the basaltic volcanics of Mt Erebus and Mt Discovery. The rifting created extension forces throughout the area weaken the crust and allowing the upwelling of the mafic intrusives. From the geological map the dykes trend NE – SW trending along the edge of the Ross Sea and Transantarctic Mountains.

The geological map has provided a better understanding of the spatial distribution of lithologies throughout the Denton Hills area.

6.2. Interpretation of Geomorphic Evolution

6.2.1. Introduction

This was the first large scale geomorphological mapping to be conducted in the Denton Hills area. Previous studies have gone into the great detail defining certain features and process, and supplementing them with smaller localised maps. The geomorphological map (Appendix Sheet 2) shows the connections between localised features and the regional geomorphic evolution.

6.2.2. Ice

Ice has been the predominant mechanism for the evolution of the valleys since the initial alluvial formation suggested by Sugden et al., (1999). The location of the valleys has allowed ice to not only flow down the valley but also flow into the valley from the mouth. On the geomorphological map (Appendix Sheet 2) ice movement has been indicated across the drift sheets. This shows the ice from McMurdo Sound reaching nearly 10 km into the Miers Valley, and significantly into the Marshall and Garwood Valleys. The map also shows the movement of ice from the western end of the valley down toward the east, however most to the glacial sediments have been removed during subsequent events. Also observed high on ridges within the area are localised small alpine glaciers or glacial deposits. These deposits have formed in a local cirque, which would have fed ice down from the ridgelines into the larger valley glaciers.

6.2.3. Lake Ice Conveyor

The geomorphological map indicates much of the Miers Valleys (and smaller portions of Marshall and Garwood Valleys) have been covered by proglacial lakes. These lakes formed when the mouths of the valleys were blocked by ice from McMurdo Sound. The ice would have prevented water escaping from the valleys, creating a lake. The lake would have been permanently covered by a layer of ice. This ice would allow the transport of sediment into the valley (Figure 4.15). Lake sediments would have formed in the base of the lake and then as the ice retreated and the lake began to drain, the sediments on the ice surface were lowered onto the lake sediments. These and also ridgeline which represent the ice layer margins, have been preserved on the floor of the valleys. Dating of the sediments suggest the last event occurred during the LGM, this is further discussed below.

6.2.4. Alluvial

On the geomorphic map the major drainage streams, which only became active in the summer months, are illustrated. The streams and rivers for most of the year are frozen, yet when the sunlight hours increase during the summer months, these systems become very active draining vast amounts of melt water through the valleys. The amount of melt water transported has created a large erosion mechanism in the valleys, craving out deep channels. The flow of water also can carry vast amounts of sediment, either transported out of the valleys into McMurdo Sound or depositing an in delta, alluvial fans or within the lakes. The drainage system has altered the landscape most in the recent history of the valleys.

6.2.5 Aeolian

Extreme winds blow through the valleys, predominantly from the west in the form of katabatic winds which flow off the plateau. The winds have both deposited and eroded features within the valley. The geomorphic map shows where features have been deposits, usually aeolian sands are found in areas, which are in the lee of prevailing winds. A good example of this is the moraines in the front of the Joyce Glacier (Garwood Valley), here the moraines have been covered in a layer of aeolian sands. The wind would flow from the west across the glacier surface, when it reaches the face there is about 20 m difference in height, this difference creates eddy depositing the sand onto the moraines.

The power of the wind to lift and move both sand and ice creates a large erosion force. High on the ridgeline the bedrock has been eroded, creating tafoni. The wind flowing through the valleys affects the floors of the valleys too, smoothing the edges of boulder into ventifacts and forming windblown sediment deposits.

6.2.6. Interpretation of the Geomorphological Evolution

The evolution of the study area can be interpreted by observing the geomorphic map, which shows the deposits left behind by the many processes, which have altered the landscape. Ice has been the foremost mechanism for the shaping of the valleys however the deposits found on the floors of the valleys indicate a more complex history. The dominant features throughout the area are proglacial lake sediments which were created when the valleys the valleys were occupied by a large proglacial lake. The lake transported sediment into the valleys creating ridges and mounds of sediment.

Presently the more active processes of alluvial and aeolian, have subsequently imprinted over the older glacial and proglacial lake deposits. The alluvial processes have either covered the floor by depositing alluvial fans or removed sediment forming channels and streams.

6.3. Cosmogenic Nuclide Dating

6.3.1. Introduction

This was the first study to use cosmogenic nuclide dating to date glacial features associated with glaciers flowing into the valleys from the west. The dating was conducted to refine the timing of events into the geomorphic evolution.

6.3.2. Garwood Glacier Moraines

Moraines surround the Garwood Glacier were tested, providing good results. As previously discussed the data suggests the inner moraine (Moraine 1) formed about 22 ka and the outer moraine (Moraine 3) formed between 38 - 47 ka. Dating these features has allowed extrapolations to be made to the surrounding landforms and surfaces. The middle moraine of the Garwood Glaciers sequence must have formed between 22 – 47 ka. Outside Moraine 3 surfaces must be older than 38 ka unless formed by more recent processes (e.g. alluvial or aeolian).

Moraine 1 (inner moraine) has an age suggesting formation during the LGM, which means the Garwood Glacier was at a greater extend during the LGM and has since retreated back to the present limit. The age helps determine the behaviour model for alpine style glaciers in the Antarctic environment, suggesting they behave in sequence with the global glacial cycles (Refer to Figure 6.1).

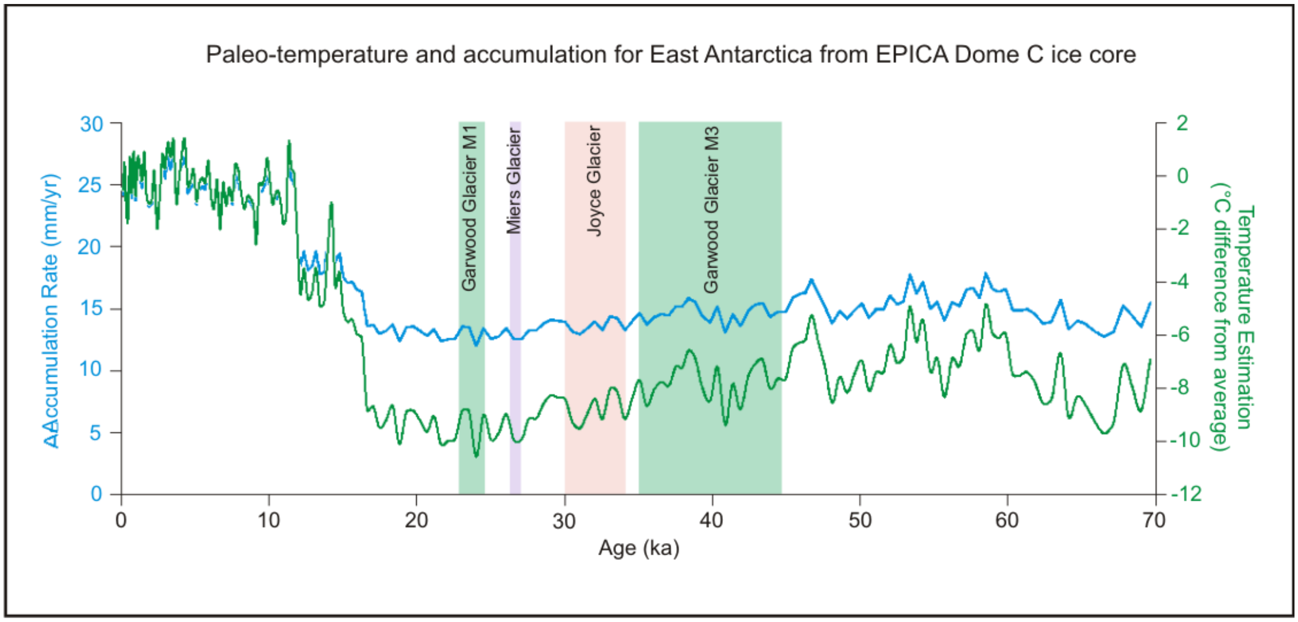

Figure 6.1: Palaeo-temperature and accumulation data collect from the EPICA Dome C ice core, East Antarctica. The SED ages from the Miers, Joyce and Garwood Glaciers are projected onto the data. There is no strong correlation between the ages of the features and a significant temperature or accumulation change (Modified from Jouzel et al., 2007).

6.3.3. Joyce Glacier Lineament

The Joyce Glacier lineament provided the best correlation between samples giving an estimated formation age of 30 ka. This lineament represents a period when the Joyce Glacier extended into the Garwood Valleys considerably further, estimated to have been about 500 m further than the current glacial limit. As the feature is a lineament along the northern valley wall, it can be extrapolated that the surfaces at higher than the lineament must be older than 30 ka. It would also suggest that the western end of the Garwood Valley has been exposed for this period of time, allowing the formation of alluvial geomorphic features.

6.3.4. Miers Valley Lineament

The lineament dated from above the Miers Glacier on Holiday Peak provided a good estimated age of 27 ka. This age would represent the last period the ice was at this height (about 320 m), leaving the lineament along the side of the peak. The age suggests the Miers Glacier was at a greater volume at the start of the LGM and has subsequently retreated. This means the Miers Valleys above 320 m has been exposed for at least 27 ka, subjected to weathering processes and being stable to allow for biological activity (e.g. lichen growth). Surfaces below 320 m have been highly modified since 27 ka this is further discussed with the samples from the Adams Glacier features.

6.3.5. Adams Glacier Ridge

The Adams Glacier feature is a ridge adjacent to Miers Lake, and has provided the most interesting dates. Although none of the samples correlate with each other, AG4 can be assumed to be the true exposure age as both ^{10}Be and ^{26}Al isotopes agree, with an exposure age of 14.5 ka. This age correlates with the end of the LGM, and when considering the method of formation would match with the end of the occupation of the proglacial lake. The lake would have formed during the LGM, as ice from McMurdo Sound would have blocked the mouth of the valley. The ridge would have formed at the end of the occupation of the lake, which would have occurred went the ice melted from mouth of the valley. It can also be inferred that the floor of the Miers Valley underwent vast changes during the presence of the lake (during the LGM), and for a short period after while the ice melted from the mouths. This would have also occurred in the other valleys.

6.3.6. Cosmogenic Dating Summary

A sequence of events can be discussed by looking at the ages prior to LGM and the other age which date to the LGM. The Miers and Joyce Glaciers samples date to the onset of the LGM, 27 ka and 30 ka respectively. The dates are just prior to the onset of the LGM suggesting the larger EAIS drainage glaciers have little to no influence with global glacial cycles. The dating however only shows that the glaciers did not expand pass this limit during the LGM, the size immediately before the LGM is unknown therefore some expansion during the LGM may have occurred. It does indicate that there was not a large expansion during the LGM unlike other place around the world (Figure 6.2 & 6.3).

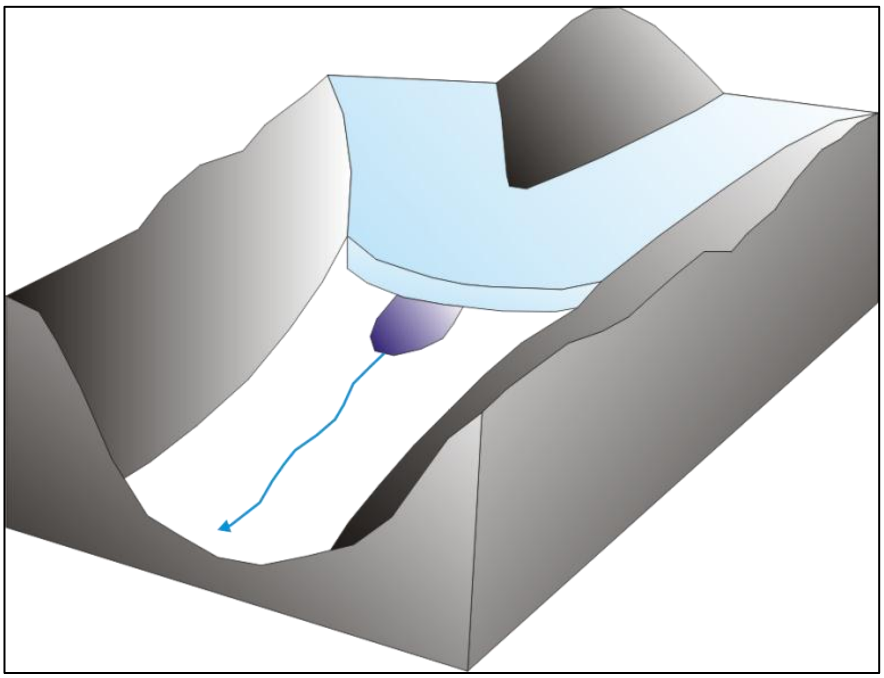

Figure 6.2: Block diagram of the Miers Valley at the onset of the LGM (approximately 27 ka). The Adams and Miers Glaciers are at a larger volume, convoluting around Holiday Peak, this represents the period which the Miers Valley lineament was formed.

The dating has also provided some good ages which directly relate to the LGM period. The inner moraine surrounding the Garwood Glacier dates directly to the LGM, with an estimated age of formation of 22 ka. The dating suggested the moraine formed about 22 ka, this glacier is an alpine style glacier being fed from a localised catchment area. This dating on the Garwood

Glacier could suggest there is a correlation between alpine glacier in Antarctica and global glacial cycles.

The SED dating has shown the mouths of the valleys were blocked during the LGM by ice in McMurdo Sound (Figure 6.3). This created proglacial lakes which filled the floors of the valleys, these lakes could transport material along their frozen surface into the valleys creating ridges and mounds. The dating of one of these ridges suggested the lakes started to drain about 14 ka after the LGM, when the ice at the mouths of the valleys would have slowly melted away. This date also correlates with and radiocarbon dates from Denton & Hughes (1981) and Hendy (2000) on algae found in the lake sediments.

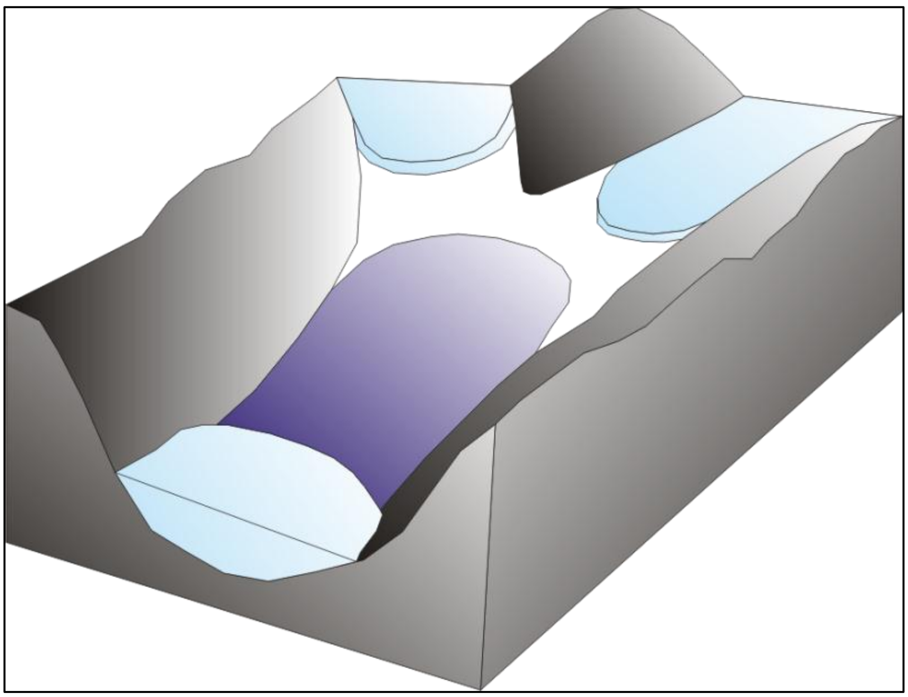

Figure 6.3: A block diagram of the Miers Valley during the LGM with the ice blocking the mouth of the lake forming a proglacial lake. This lake occupied most of the Miers Valley and would have had a permanent ice cover. During the LGM the Adams and Miers Glaciers are smaller in volume.

After the ice had melted from the mouth of the valleys the proglacial lakes would have drained depositing the remaining sediment onto the valley floor, the Adams and Miers Glaciers have stayed at below the Miers lineament, indicating a smaller volume than pre LGM (Figure 6.4).

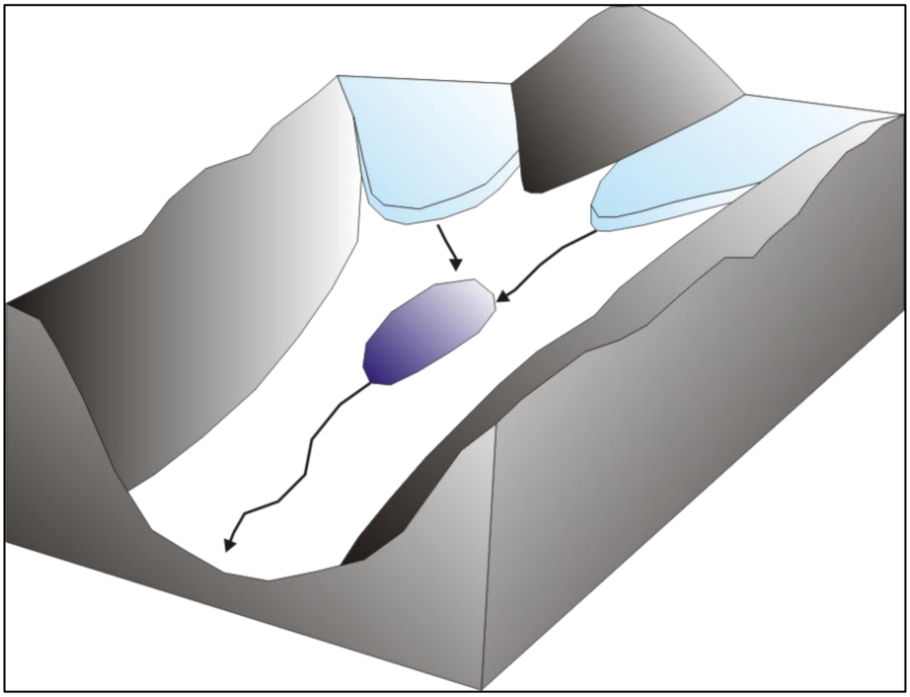

Figure 6.4: Block diagram of the Miers Valley post LGM, the ice has melted from the mouth of the valley allowing water to flow out to McMurdo Sound. This drained the lake, reducing the size down to the current level. The Adams and Miers Glaciers are still at a volume less than pre LGM.

7. Conclusions

7.1. Conclusions

This study focused on the geology and geomorphology of the Denton Hills area of the southern McMurdo Dry Valleys. The study was conducted to provide detailed mapping of both the geological units and geomorphological features, allowing interpretations to be made into the evolution of the area. Further to the mapping, surface exposure dating was conducted to allow interpretation into the timing of glacial events and surrounding surfaces.

The bedrock geology of the Denton Hills area is dominated by Pre-Cambrian to Ordovician geological units, which have been uplift during the rifting of the Ross Sea creating the Transantarctic Mountains. The oldest geological unit is the Koettlitz Group, comprised of metasediments, which are found scattered throughout the area. The group has undergone amphibolite grade (high temperature, low pressure conditions) metamorphism from original marine sediments. After the initial formation of the metasediments, granitoids were emplaced along large NW – SE plutons during Cambrian – Ordovician. The granitoids are the dominant bedrock lithology throughout the Denton Hills area, becoming more dominant to the west. The granitoids were emplaced during two different phases and two different tectonic regimes; a continental arc and an extension zone. Since the intrusion of the granitoids, there is little evidence of other geological activity, until mafic intrusions during the Cenozoic. The mafic intrusions show a reactivation of the extension zone along the Transantarctic Mountains. This reactivation is associated with the rifting of West Antarctica. The rifting stretched the crust which allowed basaltic dykes to be emplaced throughout the area in a NE – SW trend.

The geomorphic evolution of the Denton Hills area can be summarised by a series of reoccurring events. The initial formation of the area was by fluvial processes, draining water from a warmer continent. As the continent cooled the valleys started to be modified by ice, as it started to flow through the valleys. Ice eroded the valleys, deepening and widening them, creating steep valley walls and flat valley floors. The valleys floors have since been filled with more recent deposits, which indicate ice has flown from both directions (up and down the valleys). A dominant geomorphic deposit which covers a significant area of the valley floors has been created not by the ice but from proglacial lakes. These lakes would have formed when the eastern ends of the valleys were dammed by ice. The proglacial lakes would have occupied most of the valley floors altering or removing previous geomorphic features. Also these proglacial lakes have proved to be an important mechanism for the transport and deposition of sediment into the valleys. The frozen surface of the lake acts like a conveyor, moving sediment into the valleys, far beyond the ice limit.

This is the first study in the area, where combined detail mapping of both the bedrock geology and geomorphic features have been completed. The geological map has increased the detail and complexity shown previously by Blank et al., 1963, the geomorphic map is the first detailed mapping done of the whole area.

Supporting the geomorphological map in the evolution of the area, surface exposure dating was completed to allow ages to be determined. The dating method provides age constraints for the formation of the selected features, and also allows interpretations to be made on the formation of adjacent surfaces. Dating of material associated with the last proglacial lake to have formed in Miers Valley, suggested that the lake occupied the valley from the beginning of the LGM through to about 14 ka. This date can then be extrapolated to suggest that the majority of the Miers Valley has been unaltered for last 14 ka. This age correlates with the radiocarbon dates of lake carbonates and algae (Hendy, 2000) and also the SED of M I material conducted by Brook et al., (1995).

The SED dates from features associated with glaciers draining the EAIS, suggest there was little fluctuation in the size during the LGM. A field observation also noted that the Joyce and Miers Glaciers do not well-defined terminal moraines, suggesting the moraines may have been removed by the occupation of a proglacial lake associated with the expansion of the WAIS. The SED dates from the Garwood Glacier suggest there has been a series of retreats from about 45 ka, with the last feature being directly associated to the LGM. These SED ages from the Garwood Glacier could suggest that alpine style glaciers in Antarctica environment react in-phase with the global glacial cycle.

7.2. Recommendations for further research

This study provided maps of both the geological and geomorphological features throughout the Denton Hills area. While creating the maps some lithologies had to be generalised due to restrictions in both field time and analysis techniques. With greater time and equipment for analysis the maps could be refined, separating the granitoids, either into the two suites or into the individual granitoids.

Further use of surface exposure dating would improve the timing of glacial events and further correlate events to the global cycles. This research could then prove if Marchant et al., (1994) 'out-of-phase' hypothesis occurs or if all Antarctic glaciers react to the global cycle in the normal manner.

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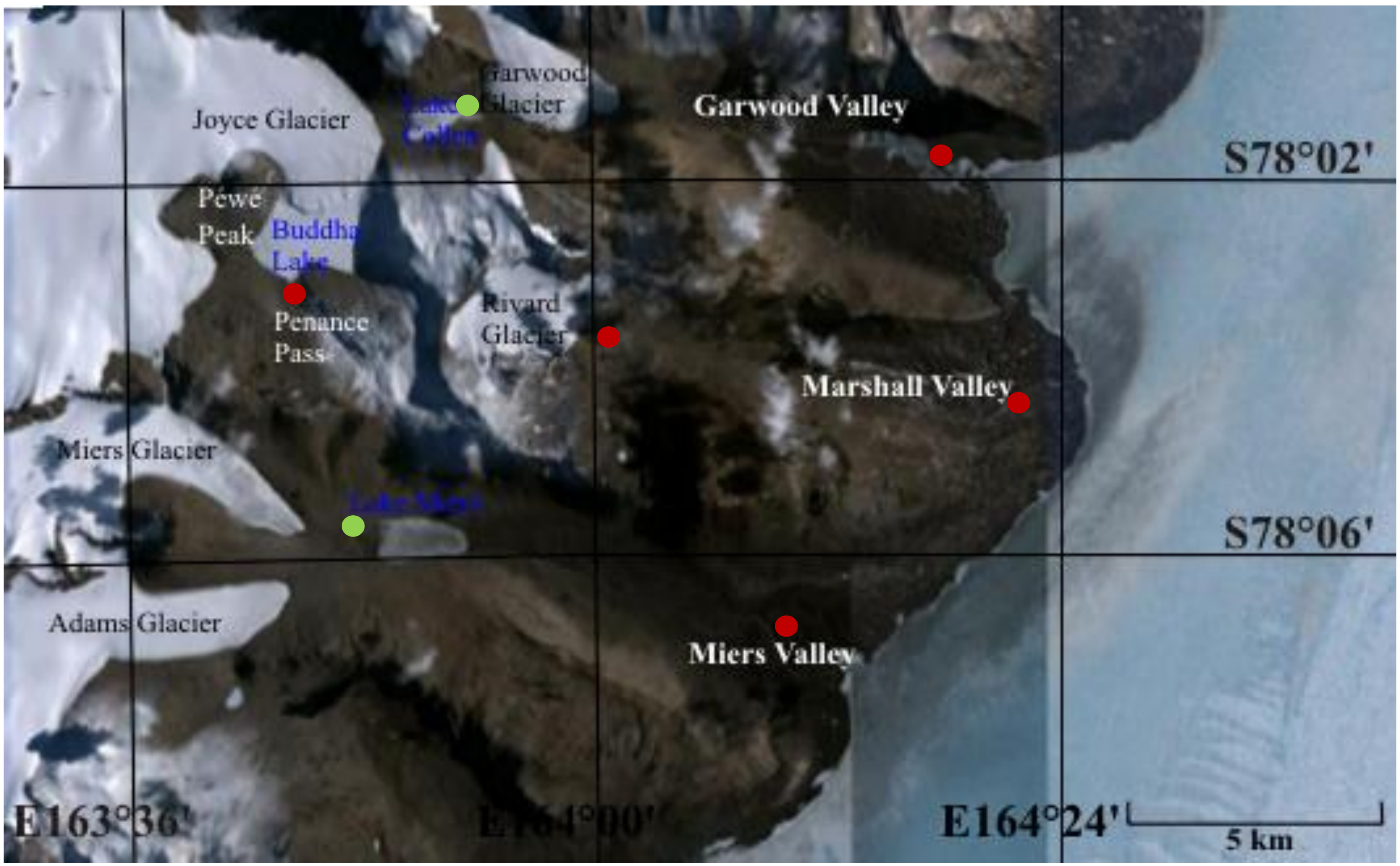
Appendix 1

Location of camps during the nzTABS project (7th November – 27th November 2008). The two green circles are the main camps and the red circles are axillary camps.



Appendix 2

Metamorphic Deformation

Deformation of the basement rocks was investigated by Fikkan, (1968), which looked into the granitoids and metasediments. By investigating the deformation structures in the metamorphic rocks and foliations within the granitoid intrusive, they were able to identify at least two generations of folding events. Fikkan, (1968) also recognised that many of the intrusive margins lacked any significant chilled margin against the Koettlitz Group metasediments. The absence of chilled margins against the metasediments, suggest that the metasediments must have been about the same temperature as the intruding granitoids. To achieve the same temperature, it requires the metasediments must have been buried well beneath the ground surface, this would imply the metasediments are also under pressure at a plastic state.

A study by Findlay (1978) suggested the geological units had undergone low pressure – high temperature metamorphism. By investigating the structural features they suggested that there was three phases of folding; F1 – mesoscopic isoclinals folds, F2 – upright angular to megascopic folds with a step-like geometry, F3 - broad open wrap with a steeply north to northwest plunging axis (Findlay, 1978). The folding within the Koettlitz Group is shown in Findlay (1978) block diagram of the Miers Valley (Figure A). Within the marble, Blank et al. (1963) had believed to have identified Archaeocyathids, Findlay (1978) suggested that what they had identified were actually quartz rods that had formed during the folding of the marble, due to the temperature.

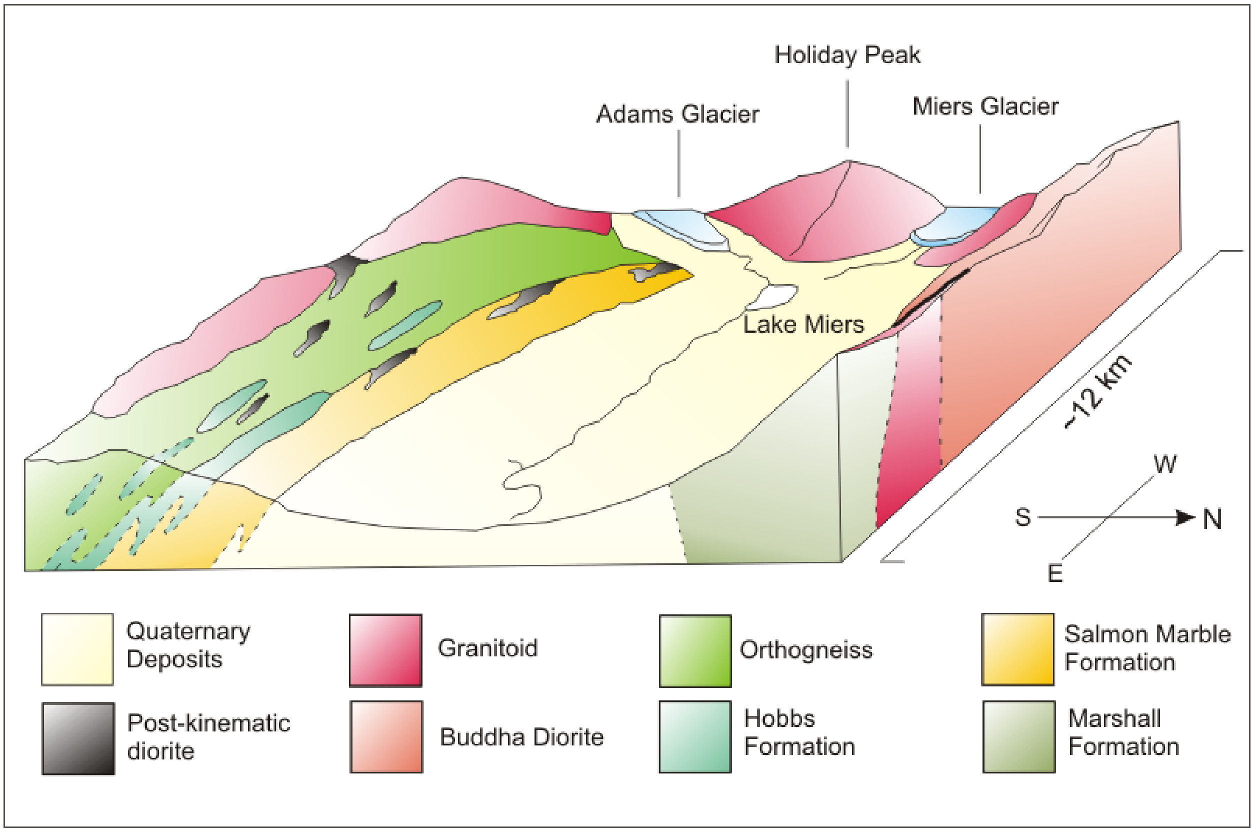


Figure A: A sketch block diagram, showing the deformation of the Koettlitz Group; Marshall Formation, Salmon Marble Formation, Hobbs Formation and Orthogneiss. Diagram modified from Finlay (1978).

Appendix 3

Scaling Factors

^{10}Be	^{26}AL				
1.388E+06	7.010E+05	y	Density =	2.70	g/cm^3
4.994E-07	9.888E-07	y^{-1}	Atten length=	150.00	cm^2/g
4.600	31.100	at/g^{-s}	Atten err =	4.0	cm^2/g

Field ID	Latitude (Decimal Degrees)	Altitude (masl)	Altitude to Antarctic Air Pressure (hPa)	Stone/ Lal Correction Factor	Sample Thickness (mm)	Sample Thickness Correction	Topographic Shielding Correction	^{10}Be Site Prod Rate (at/g/yr)	^{26}Al Site Prod Rate (at/g/yr)
GG3	-78.015	409	937.2	1.885	50	0.956	0.997	**8.269**	**55.907**
GG1	-78.015	412	936.8	1.889	30	0.973	0.996	**8.428**	**56.984**
GG4	-78.015	400	938.3	1.867	50	0.956	0.996	**8.184**	**55.329**
GG5	-78.015	420	935.8	1.904	50	0.956	0.991	**8.302**	**56.125**
GG8	-78.015	388	939.8	1.846	50	0.956	0.990	**8.038**	**54.342**
GG9	-78.015	391	939.4	1.851	40	0.965	0.992	**8.145**	**55.070**
GG10	-78.015	389	939.7	1.847	50	0.956	0.994	**8.077**	**54.605**
JG1	-78.015	390	939.5	1.849	50	0.956	0.992	**8.067**	**54.541**
JG2	-78.015	387	939.9	1.844	40	0.965	0.993	**8.125**	**54.932**
JG4	-78.1	390	939.5	1.849	50	0.956	0.993	**8.075**	**54.595**
AG3	-78.1	181	965.8	1.501	40	0.965	0.986	**6.567**	**44.396**
AG4	-78.1	184	965.4	1.506	50	0.956	0.981	**6.499**	**43.936**
AG5	-78.1	179	966.0	1.498	50	0.956	0.985	**6.495**	**43.909**
MG3	-78.1	317	948.6	1.722	50	0.956	0.997	**7.549**	**51.036**
MG4	-78.1	318	948.5	1.723	40	0.965	0.997	**7.625**	**51.555**
MG5	-78.1	324	947.8	1.734	50	0.956	0.997	**7.606**	**51.423**

Be/Al Ratio

		^{10}Be Concentrations		^{10}Be Concentrations (SLHL)	26Al Concentrations		Ratios	
Cathode #	**Field ID**	**^{10}Be Conc (atoms/g-Q) (1E6)**	**^{10}Be Conc Err (atoms/g-Q) (1E6)**	**^{10}Be Conc (atoms/g-Q) SLHL**	**^{26}Al Conc (atoms/g-Q) (1E6)**	**^{26}Al Conc (atoms/g-Q) (1E6)**	**Al/Be Ratio**	**Al/Be Ratio Error (abs)**
B3773	GG3	0.3651	0.0088	203117	3.2718	0.7111	8.960	1.9476
B3772	GG1	0.2934	0.0072	160128	2.4326	0.2927	8.291	0.9976
B3887	GG4	0.2016	0.0085	113315	0.9896	0.3628	4.909	1.7997
B3888	GG5	0.2940	0.0092	162906	1.3404	0.3701	4.559	1.2589
B3774	GG8	0.4836	0.0121	276761	3.3902	0.4064	7.010	0.8403
B3889	GG9	0.1932	0.0074	109094	1.5351	0.2013	7.947	1.0420
B3775	GG10	0.1785	0.0075	101677	1.2327	0.2819	6.905	1.5793
B3890	JG1	0.2530	0.0091	144261	1.7520	0.3583	6.925	1.4161
B3776	JG2	0.2349	0.0067	132981	1.5945	0.2487	6.788	1.0588
B3891	JG4	0.2502	0.0109	142545	2.0858	0.7024	8.336	2.8071
B3777	AG3	0.1281	0.0040	89760	0.4964	0.1618	3.874	1.2628
B3892	AG4	0.0934	0.0040	66082	0.6386	0.1513	6.841	1.6206
B3778	AG5	0.1502	0.0044	106390	1.3852	0.2626	9.222	1.7482
B3893	MG3	0.2776	0.0098	169182	2.0739	0.3845	7.470	1.3850
B3894	MG4	0.2063	0.0114	124466	1.5348	0.2983	7.439	1.4457
B3895	MG5	0.2055	0.0082	124308	1.3572	0.3015	6.603	1.4668